INFLUENCE OF VOLATILES IN THE UPPER MANTLE ON THE DYNAMICS OF THERMAL THINNING OF THE LITHOSPHERE

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ABSTRACT

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Areas of Cenozoic tectonic and magmatic activity (including high plateaus and continental rifts) are characterized by anomalous low-velocity and low-density zones in the upper mantle. Petrochemical studies of Cenozoic volcanism have revealed such changes of magma composition with time that prove a consecutive uplift of magma sources. Thus, it is supposed that the process of tectonic rejuvenation is caused by an ascent of anomalous mantle to the base of the lithosphere, resulting in partial melting of lithospheric material, its consecutive replacement by material from the anomalous mantle, lithosphere thinning, and, hence, isostatic uplift of lithospheric blocks. A model of thermal thinning of the lithosphere that is specified by a 1-D heat-conductivity problem for the lithosphere with a moving lower boundary, is proposed as a model of Cenozoic tectonic activation. The presence of H_2O and CO_2 fluids in the upper mantle is taken into account. Numerical modelling of the process has revealed that the composition of the upper mantle and fluid phase has a strong influence on the dynamics of the process. The presence of volatiles in the upper mantle leads to the appearence of maxima and minima on the solidus curves. Lamination of fusible and refractory layers in the upper mantle may lead to a sharp change in lithosphere-thinning velocities and, hence, to a discrete character of surface vertical motions. The thickness of the lithosphere in a new equilibrium position is calculated for a different composition of the upper mantle and different values of heat flow supplied to the base of the lithosphere; the results show that for the models that seem best to fit present knowledge of the upper mantle composition, melting of the lower crust may take place only in the later stages of the process, when the lithosphere-anomalous mantle boundary approaches its new equilibrium position.

INTRODUCTION

Some areas of the Earth that for a long time developed as platforms have undergone an intensive tectonic activation in the Cenozoic. The result of such activations was the formation of large mountain systems (like Tien Shan), high plateaus (e.g., the Colorado Plateau), and continental rifts (e.g., the Baikal and East African rifts).

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The areas of Cenozoic activation are characterized by some general features of their deep structure. Gravity studies show that on a regional scale they remain in approximate isostatic equilibrium (e.g., Kaula, 1972; Zorin and Florensov, 1979), the level of isostatic compensation being largely in the upper mantle (Nersesov *et al.*, 1975; Thompson and Zoback, 1979). Anomalous mantle of a low density, the existence of which is proved by large negative Bouguer anomalies (about -150 -250 mGal for different areas of regional uplift) (e.g., Woollard, 1972; Girdler, 1975) and by long-wavelength isostatic and free-air gravity anomalies, plays a certain role in the realization of isostasy (Artemiev, 1966; Zorin and Florensov, 1979).

The upper mantle of the areas of Cenozoic tectonic activation is characterized by low seismic velocities and high attenuation (Fairhead and Girdler, 1972; Illies *et al.*, 1979; Keller *et al.*, 1979; Vinnik and Saipbekova, 1984). In some areas (e.g., high mountain regions of Central Asia and Southern Siberia, the western United States) a good negative correlation between the average seismic velocities in the upper mantle and surface relief has been noted. Lithosphere blocks with low seismic velocities in the mantle coincide with areas of large negative Bouguer anomalies, close to zero isostatic gravity anomalies, and with the areas of increased heat flow.

The lithosphere of the areas of Cenozoic regional uplifts is 20–50% thinner than before the initiation of the tectonic process (about 25–30 Ma ago) (Crough and Thompson, 1976; Illies *et al.*, 1979; Spohn and Schubert, 1982; Zorin and Osokina, 1984), while in some regions the crust beneath the uplifted areas is thickened, mainly due to the growth of the "basalt" layer. For example, thickness of the "basalt" layer beneath mountain ridges of Northern Tien Shan is 40 km, while that of the "granite" layer is only 10 km. The corresponding thicknesses beneath depressions of the same area are 25 and 20 km (Belousov, 1964).

The average velocities of lithosphere thinning caused by the anomalous mantle uplift are 2.00 km/Ma for the Colorado Plateau, 2.40 km/Ma for Eastern Tien Shan and the Rhenish Massif, 2.60 km/Ma for the axial part of the East African rift, and 2.67 km/Ma for the Baikal rift zone.

MAGNETISM OF THE AREAS OF CENOZOIC TECTONIC ACTIVATION

The areas of regional uplift are characterized by Cenozoic, mainly basaltic, volcanism. Petrological and geophysical investigations indicate a consecutive uplift of magma sources in these areas and make it possible to recognize some evolutionary trends by comparing data for different structures. Studies of Cenozoic volcanism of the Kenya rift, the Rhenish Massif, Central Tien Shan and other areas, have revealed consequence in changes of magma composition with time from the deepest ultra-alkaline magmas (melanephelinites, melaleucitites, and derived carbonatites originally generated at 80–100 km) and alkali olivine basalts (generated at 60–80 km depth), to low-Ca phonolites and trachytes (generated correspondingly at 40–50 and 35–40 km depth) for the Kenya rift (Ringwood, 1975; Yoder, 1976).

Laboratory studies of the physical and chemical conditions of magma generation in multicomponent silicate systems under adiabatic decompression explain the nature of the regularity described above and give a petrological basis for the model of anomalous mantle uplift and a process of lithosphere thinning. This model is reinforced by geochemical data on the temporal decrease of magma alkalinity and of the concentration of incompatible and trace elements (Grachev, 1977; Wendlandt and Morgan, 1982). A mantle source of alkali magmatism is enriched in rare lithophile elements (Rb, Cs, Ba, Nb, LREE) and radiogenic isotopes. This is an indication that anomalous mantle rises from a deep undepleted geochemical reservoir. The question as to a what depth large-scale mantle asthenoliths are generated is still unclear. However, nowadays it seems more likely that they are related to a boundary layer separating the upper and lower mantle, convecting more or less independently (Nakanishi and Anderson, 1982; Anderson, 1987).

THERMAL THINNING OR GRAVITY INSTABILITY?

The process of formation of intracontinental areas of regional uplift has been the focus of a number of recent investigations. According to Crough and Thompson (1976), Mareschal (1983b), Neugebauer (1983) and Spohn and Schubert (1983), conversion of a stable platform to a region of tectonic and magmatic activity (including orogens and continental rifts) is connected with a consecutive displacement of the front of a convective heat source, the development of deep diapirism, and uplift of anomalous mantle material to the base of the lithosphere. Petrological data, for instance for the Rhenish Massif, reinforce a progressive uplift of a source of magmatic material with a velocity of 5–10 km/Ma (Neugebauer *et al.*, 1983).

Interaction of an uprising anomalous mantle with the lithosphere can be modelled by lithospheric thermal thinning (Crough and Thompson, 1976; Spohn and Schubert, 1982, 1983; Gliko *et al.*, 1985) or by Rayleigh-Taylor instability (Mareschal, 1983a; Neugebauer, 1983). The dominance of one or the other mechanism is governed by the thermo-mechanical state of the anomalous mantle material, the lithospheric rheology, and perhaps by some other physical and chemical processes.

If a continental lithosphere is in unstable mechanical equilibrium because of a density contrast between its lower part and underlying asthenosphere, a gravity instability develops. This process leads to a diapiric rise of the asthenosphere, detachment of the lower lithosphere and its displacement by a hot material of anomalous mantle. The data on the amplitudes and velocities of Cenozoic uplift of the western United States are in good enough agreement with the result of numerical modelling of a gravity instability development (e.g., Bird, 1979).

The mechanism of a gravity instability can provide the required rates of lithosphere thinning only when the viscosity of the lithosphere does not exeed 10^{23} poise (Neugebauer, 1983). In contrast, the process of thermal thinning of the lithosphere plays a significant role when the lithosphere viscosity is greater than 10^{23} poise, as is typical for the cold lithosphere of platform areas. Thus, the mechanism of lithospheric thermal thinning is acceptable for such areas as, for instance, the Tien Shan and the East-African rift, that have developed as platforms before the initiation of the tectonic process.

In this paper we represent some results of the study of the lithospheric thermal-thinning process. Two different variants of a formulation of the problem are acceptable. In the model of "convective thinning" it is supposed that the lower lithophere material is heated to the anomalous mantle temperature, exceeding the solidus by heat convected to the base of the lithosphere. Erosion of the lower lithosphere causes lithospheric thinning and isostatic uplift of the lithospheric blocks. However, it is a slow enough process, and it takes 30–35 Ma to form an asthenospheric dome only if heat flow from the anomalous mantle is about 200–250 mWm^{-2} (Spohn and Schubert, 1982).

In the second model it is assumed that uplift of the anomalous mantle implies the replacement of material of the lithosphere by anomalous mantle material when the lithosphere is heated to a temperature close to the solidus and a sharp decrease of lithospheric viscosity occurs (Gliko *et al.*, 1985). Another variant of this model is that of Zorin *et al.* (1984) analogeous to R. Daly's mechanism of "large magmatic stoping". However, a mathematical formulation of the problem for these two variants is essentially identical. In these models, the heat flow from the anomalous mantle required to cause the observed lithospheric thinning is essentially less than in the model of "convective thinning".

However, the simpliest form, i.e., of a linear relationship of the solidus temperature with depth typical for "dry" systems, was assumed in all recent models. The results obtained by Jacoby (1984) show that a such formulation of a model is a serious simplification of a real process. In some areas (e.g., the Rhenish Massif), a negligible change in $\Delta\rho$ in the upper mantle, accompanied by a large change in ΔV_p, cannot be explained only by a temperature increase due to a presense of an anomalous mantle lens; it is necessary to assume the existence of volatiles and/or fluid-filled cracks in the upper mantle of this region to explain this phenomenon (Jacoby, 1984).

The purpose of the present work is a study of the influence of volatiles in the upper mantle on the dynamics of lithospheric thinning, implying a non-linear relationship between depth and the solidus temperature.

SOLIDUS TEMPERATURES OF THE SYSTEM WITH VOLATILES

Geochemical data on the composition of volcanic gases and fluids in minerals of magmatic rocks and in the upper mantle xenoliths show that a significant amount of volatiles, mainly H_2O and CO_2, exists in the upper mantle (Wyllie, 1977; Kadik, 1975) and strongly influences the melting conditions. Experimental solidus curves of silicate systems with H_2O and CO_2 are characterized by temperature maxima and minima (Kushiro *et al.*, 1968; Green and Ringwood, 1968) that must not be ignored in studies of the dynamics of lithospheric thinning.

A large amount of water in the upper mantle leads to a decrease of the

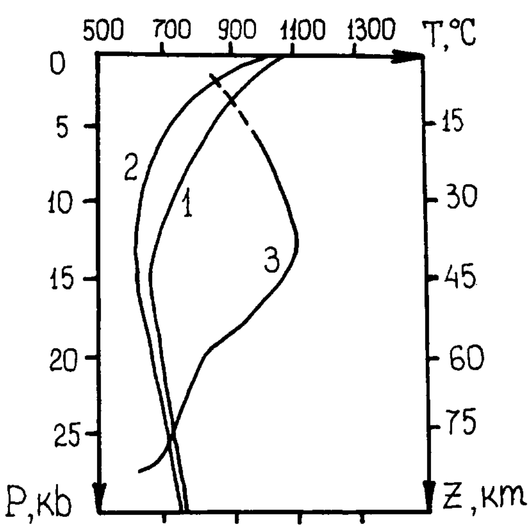

Fig. 1. The solidus versus depth for: 1) basalt + H_2O + CO_2 [1], 2) basalt with an excess of water [1, 2], 3) basalt with a small fraction of water, $H_2O < 0.5$ weigth% [2]. References: [1] Hill and Boettcher, 1970; [2] Lambert and Wyllie, 1970.

solidus temperature U_m with the growth of pressure, the maximum values of U_m being reached at $P = 15$–25 Kb. In the case of low H_2O concentration in a system ($H_2O < 0.1$–0.5 weight %), the solidus temperature increases with pressure and has a minimum at $P \approx 15$ Kb. The presence of carbon dioxide slightly decreases U_m when $P < 25$ Kb (Fig. 1).

If there exists a mixture of volatiles in the upper mantle, then a rise in CO_2 concentration leads to an increase of the solidus temperature of silicate systems as well as of rocks of basalt and peridotite composition (under a constant pressure, $P < 25$–30 Kb) (Fig. 1). Thus, a correlation between H_2O and CO_2 in the upper mantle regulates the solidus temperature and composition of derived magmas.

Apart from water and carbon dioxide, there may be present H_2, CO, CH_4, and C in a fluid. It seems very likely that the existence of these volatiles in a fluid will lead to an increase of the solidus temperature (Kadik, 1975). The most important factors that govern the content of H_2O, CO_2, H_2, CH_4, and CO in a fluid phase are the existence of free carbon and the oxygen fugacity.

In this work, we have studied the characteristic features of the lithospheric thermal thinning process for different fluid regimes. We

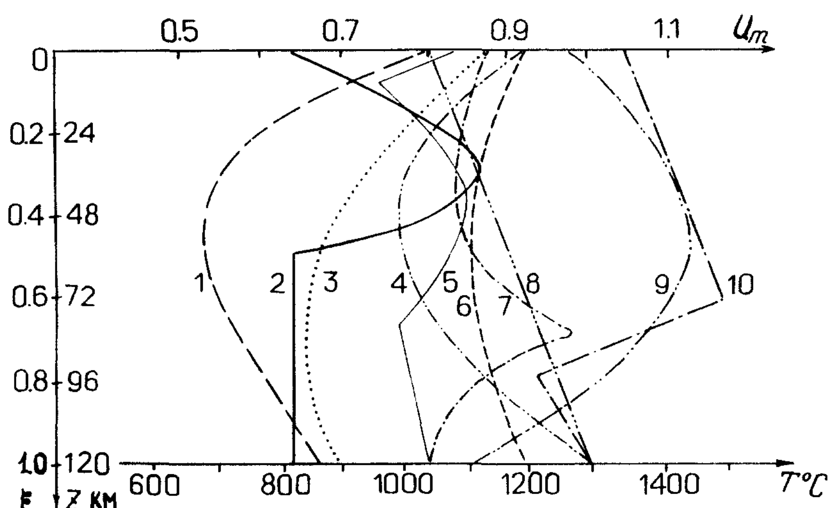

Fig. 2. The solidus versus depth for different upper-mantle composition: 1) basalt + H_2O + CO_2 [1]; 2) basalt + H_2O, $H_2O < 0.5$ weigth % [2]; 3, 6, 7, and 9) peridotite – C – H – O, the ratio $H_2O/H_2O + CO_2$ is correspondently equal to 1.00, 0.05, 0.20, and 0.00 [3, 5, 6]; 5) peridotite + H_2O, $H_2O < 0.5$ weigth % [3]; 4) ultramafic rocks + H_2O [4]; 8) "dry" solidus; 10) CaO – MgO – SiO_2 – CO_2, $CO_2 < 2$ weigth % [3]. References: [1] Hill and Boettcher, 1970; [2] Lambert and Wyllie, 1970; [3] Kadik and Frenkel, 1982; [4] Kushiro *et al.*, 1968; [5] Wyllie, 1977; [6] Wyllie, 1986.

have assumed the following models of upper mantle composition: peridotite + H_2O + CO_2 with varying concentration of water and carbon dioxide in the fluid; peridotite and basalt with low concentration of water ($H_2O < 0.5$ weight %); peridotite with an excess of water; the silicate system $CaO - MgO - SiO_2 - CO_2$ with a low concentration of carbon dioxide; basalt + H_2O + CO_2. The solidus temperatures of these systems are shown in Fig. 2.

At this stage we have assumed that the composition of a fluid phase does not change with depth; this is certainly a serious simplification. Essential changes in correlation between H_2O, CO_2, CO, H_2, and CH_4 in the fluid of an ascending mantle material (with the presence of graphite) must take place, leading to a vertical inhomogeneity of the fluid phase composition. The existence of such inhomogeneity may strongly influence the solidus temperature, and may cause an accerelation or a slowing-down of an uplift of the boundary between the lithosphere and the anomalous mantle. It seems to be important in the future to use geochemical data concerning the change of the fluid phase composition during the uplift of anomalous mantle and to obtain the corresponding solutions for the evolution of the thermal regime of the lithosphere.

MATHEMATICAL FORMULATION OF THE PROBLEM

In the frame of our model, the lithospheric structure in areas of epiplatform orogenesis is closely connected with the replacement of the lower parts of the lithosphere by hot material of anomalous mantle when the lithosphere is heated to some characteristic temperature (close to the solidus) which corresponds to a sharp decrease in lithosphere viscosity. To reduce the mathematical complexity, the lithosphere is considered as a homogeneous horizontal layer, and one-dimensional case is treated.

It is assumed that before the beginning of the tectonic process, the temperature distribution in the lithosphere $U_0(z)$ is steady-state (z-axis downwards). The initial thickness of the lithosphere, ξ_0, depends on the heat-production rate in the crustal rocks, the steady-state heat-flow value, Q_0, at the base of the lithosphere, and the thermal conductivity structure. The activation process is initiated by a sharp increase of heat flow $Q(t)$ supplied directly to the moving lower boundary of the lithosphere; thus the steady-state conditions are terminated.

Thinning of the lithosphere can be determined as the solution of an non-linear boundary heat-conductivity problem of Stefan type:

$$\frac{\partial U}{\partial \tau}=\frac{\partial^2 U}{\partial z^2}+P(z), \quad 0<z<\xi(\tau)<1$$

$$U(z,0)=U_0(z)$$

$$U(0,\tau)=0 \tag{1}$$

$$U(\xi(\tau),\tau)=1$$

$$-\left.\frac{\partial U}{\partial z}\right|_{z=\xi(\tau)}+q(\tau)a=-\alpha\frac{d\xi}{d\tau}+\beta\xi(\tau)\frac{d\xi}{d\tau}$$

where $U(z,\tau)$ is the temperature distribution in the lithosphere, $\xi(\tau)$ is the depth of the anomalous mantle-lithosphere boundary. All parameters in eqs. (1) are non-dimensional. The scaling factors of depth, temperature, and time are respectively equal to ξ_0 (the initial steady-state position of the lithosphere-anomalous mantle boundary), T_m (solidus temperature at the depth ξ_0), and $\tau_0=\xi_0^2/\chi$ (the characteristic time of lithosphere heating, where χ is thermal diffusivity). The parameter $\alpha=\lambda/cT_m$ is the Stefan number (λ is a specific heat, c is a heat capacity). Parameter $\beta=\Delta\rho g\xi_0/\rho cT_m$ characterizes the energy required for the uplift of lithospheric blocks; $q(\tau)=Q(\tau)/Q_0$ is the dimensionless heat flow at the base of the lithosphere (Q_0 is the steady-state value of the heat flow that characterizes the mantle heat flow of platform areas);

$$a=\frac{\partial U_0/\partial z}{T_m/\xi_0}$$

is a dimensionless temperature gradient in the upper mantle.

The last two equations represent the conditions on the moving boundary. The first one postulates that the temperature at the base of the lithosphere is equal to the solidus at each given depth. The second one represents the energy balance at the lithosphere-anomalous mantle boundary: the heat supplied to the base of the lithosphere results in heating up the lowermost lithosphere to the solidus temperature and its partial melting. The changes of potential energy of lithospheric blocks related to thermal expansion, partial melting, and uplift of lithospheric blocks have also been taken into account.

NUMERICAL SOLUTION AND PARAMETERS OF THE MODEL

For the numerical solution, the initial heat conductivity problem (1) is reduced to a problem for

$$v(z,\tau)\equiv U(z,\tau)-U_0(z),$$

where $v(z, \tau)$ is the deflection of the dimensionless temperature in the lithosphere from the steady-state temperature distribution $U_0(z)$:

$$\begin{aligned}
&\frac{\partial v}{\partial \tau}=\frac{\partial^2 v}{\partial z^2}, \qquad 0<z<\xi(\tau)<1\\
&\xi(0)=1\\
&v(0,\tau)=0\\
&v(z,0)=0\\
&v(\xi(\tau),\tau)=U_m(\xi(\tau))-U_0(\xi(\tau))\\
&\left.\frac{\partial v}{\partial z}\right|_{z=\xi(\tau)}+\left.\frac{\partial U_0}{\partial z}\right|_{z=\xi(\tau)}-aq(\tau)=\alpha\frac{d\xi}{d\tau}-\beta\xi(\tau)\frac{d\xi}{d\tau}
\end{aligned} \tag{2}$$

The steady-state distribution of temperature in the lithosphere is specified in such a way that $U_0(z)$ is equal to the solidus $U_m(z)$ at the lithosphere-anomalous mantle boundary when $z=\xi_0$ at time $\tau=0$, $U_0(z)$ being exponential in the crust and linear below.

The influence of mantle volatiles on the process of lithospheric thinning is specified by the choice of the function $U_m(\xi(\tau))$ that represents the change in the solidus temperature with depth for a given upper-mantle composition (Fig. 2).

The boundary problem (2) is solved numerically by a modified Bubnov-Galerkin method (Melamed, 1958; Artemieva and Gliko, 1986). The essence of the method is to reduce the initial boundary problem to an infinite system of orinary differential equations and to solve a finite system of sufficiently large dimension numerically. It is the merit of this method that one may obtain solutions with any given accuracy which depends only upon the dimension n of the finite system of diffential equations (Fig. 3). This is based on the convergence theorem which states that the solution of finite series approaches the exact solution of the problem with $n \to \infty$. However, when integrating specific systems, it is necessary to use special numerical methods to overcome difficulties connected with the rigidity of these systems.

In the numerical calculations, we assumed that at time $\tau=0$ the heat flow from the anomalous mantle is instantly increased by a factor of g which then remains constant. The calculations were carried out for $q=2$, 3.5, 5, 7, 9. The initial thickness of the lithosphere ξ_0 was taken to be 120 km, which is typical for platform areas. Other parameters were assumed as follows: thermal diffusivity $\chi=8\times10^{-7}$ m^2/s, implying a characteristic heating time of the lithosphere of $\tau_0 \approx 570$ Ma; heat capacity $c=1400$ J/kg K; specific heat of the lower lithosphere $\lambda=4.2\times10^3$ and 4.2×10^4 J/kg (this

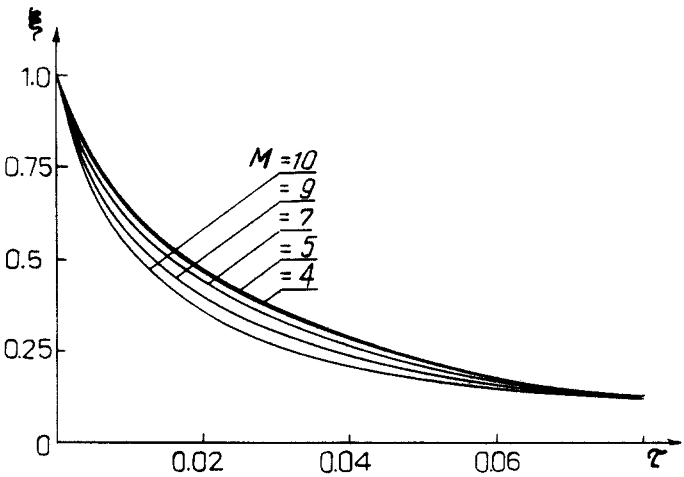

Fig. 3. Convergance of a solution of a finite system of differential equations, where M is a number of equations. Figure illustrates a lithosphere thickness (in units of ξ_0) versus time from initiation of tectonic process (in units of $\tau_0 = \xi_0^2/\chi$) for "dry" solidus when $q = 5$ and 10% melt of lithospheric material occurs.

corresponds approximately to 1 and 10% of melt of lithospheric material); density of lower lithosphere $\rho = 3.4\ \mathrm{g/cm^3}$, density contrast at the lithosphere-anomalous mantle boundary $\Delta\rho = 0.1\ \mathrm{g/cm^3}$.

RESULTS

As the first step, we have calculated the lithospheric thickness in a new steady-state position ξ^* as a function of the basal heat-flow increase $q(\tau) = Q(\tau)/Q_0$, assuming that $q = \text{const}$ and $\tau \to \infty$. From a second boundary condition of problem (2) for a new equilibrium position and taking into account that in the limit $\tau \to \infty$ a deflection of temperature from the steady-state distribution $v(z, \tau)$ must be a linear function of z, we may calculate a thickness of the lithosphere in a new steady-state position ξ^* as a function of the solidus temperature at depth ξ^* and the basal heat flow increase q:

$$\xi^* = \frac{U_m(\xi^*) - U_0(\xi^*)}{a(q-1)} \tag{3}$$

The results obtained for the solidus curves from Fig. 2 are shown in Fig. 4. One may see that the lithospheric thickness in the new steady-state position depends slightly on the type of the solidus-depth distribution if q

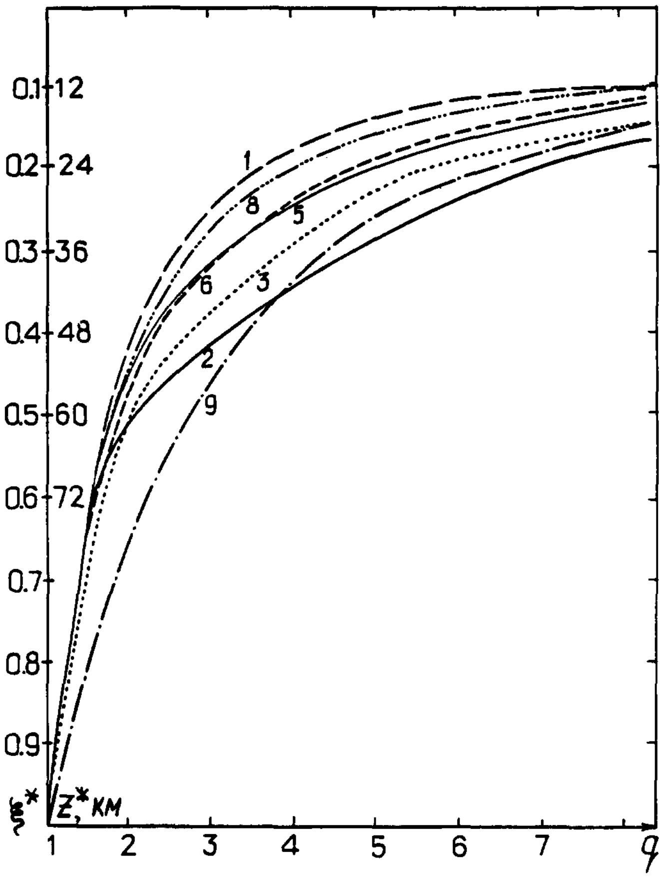

Fig. 4. Lithosphere thickness in a new equilibrium position as a function of a dimensionless basal heat flow q. Numbers of the curves as in Fig. 2.

is large and to a great extent it is determined by the type of $U_m(z)$ for relatively low values of q ($q < 5$).

When the basal heat flow increases by a factor $q = 3$–5, a fast lithospheric thinning occurs for the system basalt + H_2O + CO_2 ; but it is likely that such a composition is not typical for the real upper-mantle material. For a more realistic composition, a thickness of the lithosphere in the new steady-state position is about 40% of its initial value for $q = 3$ and about 25% for $q = 5$. This means that if the lithosphere thickness on the preactivation stage is equal to 120 km and the heat flow supplied to the base of the lithosphere increases by a factor of 3–5 from its initial value of about 20 mWm^{-2}, then during the tectonic activation process the lithosphere may be thinned to about 30–50 km, and on the latest stages of the process the lowermost crust may be melted. However, 25 Ma after the initiation of the tectonic process

TABLE I

Time required for the lithosphere-anomalous mantle boundary to approach its new equilibrium position; maximal and minimal lithosphere thickness and corresponding isostatic uplift after 25 Ma from the initiation of tectonic process for various basal heat flows.

q	τ^*	t^*, Ma	ξ_{min}, km	ξ_{max}, km	H, km
2	0.35	200	82	113	0.20–1.12
3.5	0.20	114	54	102	0.53–1.94
5	0.14	80	38	92	0.82–2.41
7	0.10	57	26	82	1.12–2.76
9	0.07	40	17	74	1.35–3.03

Here minimal lithosphere thickness corresponds to basalt + H_2O + CO_2, maximal to peridotite + CO_2; isostatic uplift was calculated for lithosphere density $\rho = 3.4$ g cm^{-3} and lithosphere-asthenosphere density contrast $\Delta\rho = 0.1$ g cm^{-3}.

(which is a typical age of Cenozoic high plateaus and continental rifts), the boundary between the lithosphere and the anomalous mantle is far from its equilibrium position (Table 1). The contemporary lithosphere thickness in the areas of regional uplift varies, on the average, from 40 to 80 km; an ascent of anomalous mantle to the base of the crust and its injection into the crust is generally observed in the zones of deep faults.

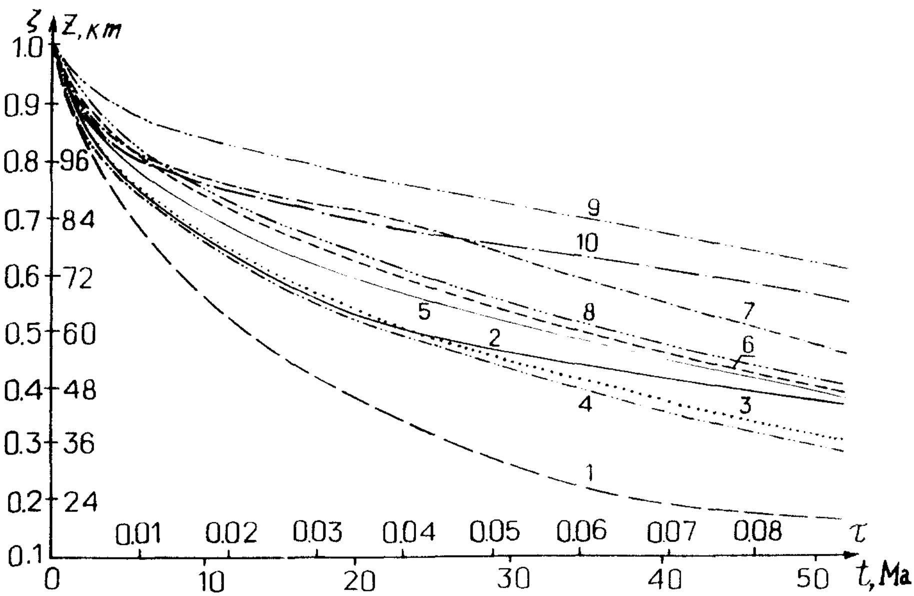

Fig. 5. Lithosphere thickness versus time passed from the initiation of the tectonic process; $q = 5$, 10% of melt of lithospheric material.

According to numerical modelling of lithospheric thermal thinning, the presence of volatiles in the upper mantle leads to the following main consequences (Fig. 5): the presence of a large amount of water in the upper mantle (curves 1, 3, 4, and 6 in Fig. 2) results in rapid lithospheric thinning in the region of low solidus temperatures, while the presence of CO^2 or a very small fraction of H_2O (curves 2, 5, 9, and 10 in Fig. 2) leads to a sharp slowing-down of lithosphere melting. Fig. 6 illustrates the influence of the heat flow from the anomalous mantle supplied to the base of the lithosphere on the dynamics of lithospheric thinning for various upper-mantle compositions.

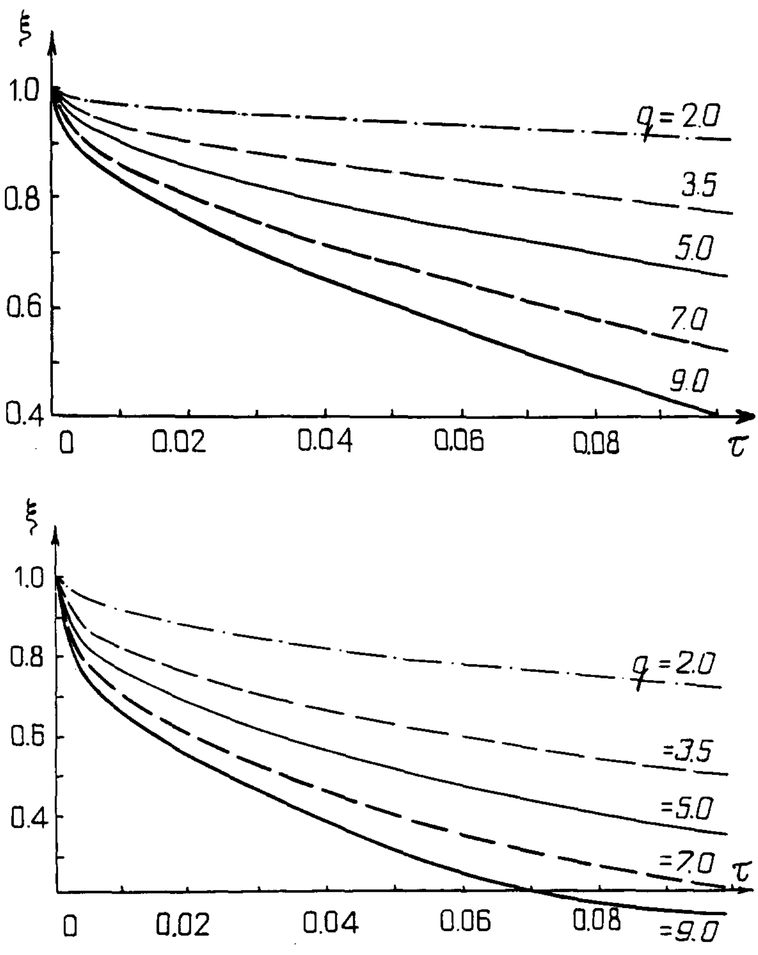

Fig. 6. Evolution of lithosphere thickness for different solidus conditions and different basal heat flow q for 10% of melt of lithospheric material. Upper figure: peridotite + CO_2; lower figure: peridotite + H_2O.

The calculations were carried out for the cases of 10 and 1% of melt in the lower lithospheric material. The change of degree of melting connected with adiabatic decompression was not considered. According to our results, the lithospheric thickness is, in general, 10–20% greater in models with 1% melt than in those with 10% melt. With an increase of q, the difference between the results obtained for different degrees of partial melting becomes smaller, and it does not exceed 5% for $q = 7$ (Fig. 7).

The most important numerical result is that the process of high-plateau formation is strongly influenced by the presence of volatiles in the upper mantle and that is expressed by a sharp change of the upwelling velocities of anomalous mantle in the region of nonliniarity of the solidus temperature with depth (Fig. 8). The results obtained for the basalt and peridotite system with different fractions of volatiles are represented, respectively, in Figs. 9 and 10 (curves 1, 2, 5, and 9 in Fig. 2). As can be seen, the rate of uprise of anomalous mantle may decrease by a factor of 2.5–3 at the time of about 1–2 Ma when a region of high solidus temperatures is reached. Such an abrupt change in the rate of uprise of anomalous mantle must result in an equally abrupt slowing-down of crustal vertical motions; in the framework of the model they are controlled by isostasy where the lithospheric blocks rise due to their thinning. In an analogous fashion, the

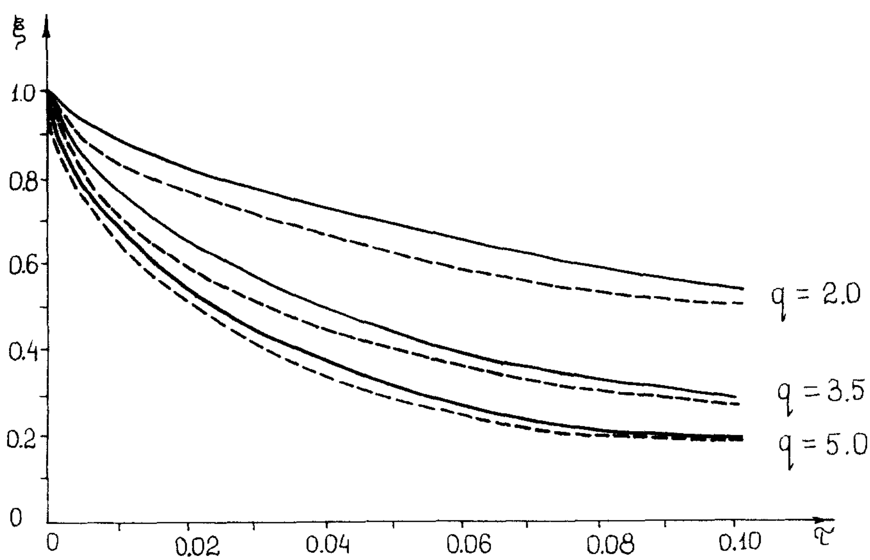

Fig. 7. Dimensionless lithosphere thickness versus dimensionless time for basalt + $H_2O + CO_2$ for different basal heat flow q. Dashed lines correspond to 10% of melt, solid lines — to 1% of melt of lithospheric material.

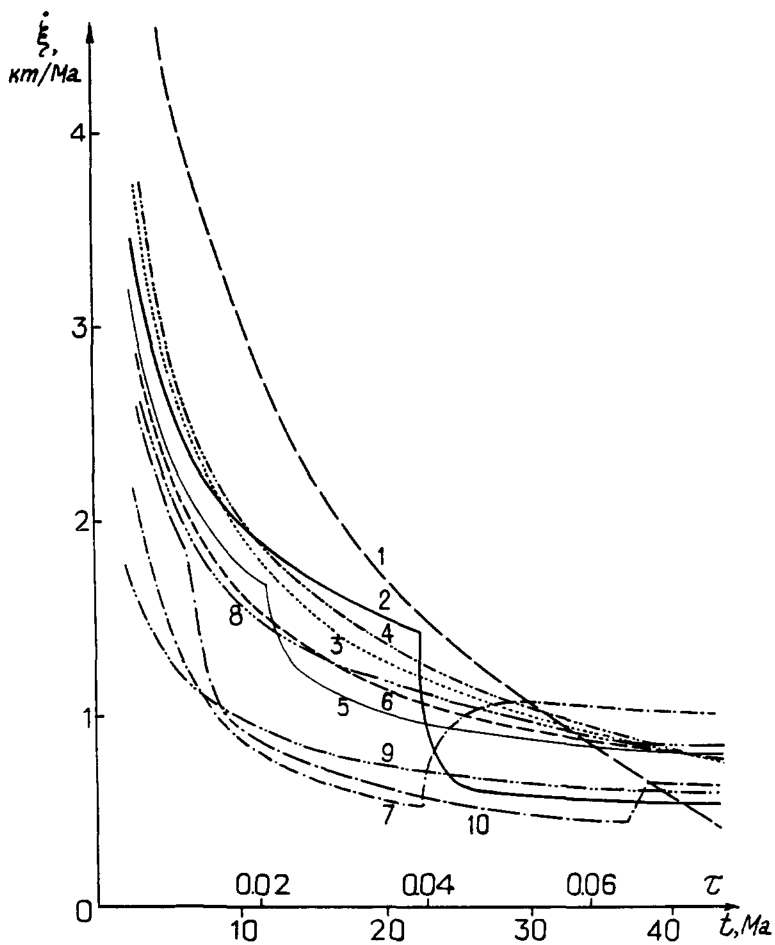

Fig. 8. Velocity of lithosphere thinning versus time from the initiation of tectonic process for $q = 5$, 10% of melt. Numbers of the curves as in Fig. 2.

rise of the boundary between lithosphere and the anomalous mantle into regions of low solidus temperatures leads to an abrupt increase of the rate of rise of the boundary and corresponding crustal uplift.

An ascent of anomalous hot mantle to the base of the lithosphere causes a gradual heating of lithospheric material. When q is constant, the velocity of heat-front propagation depends on the upper-mantle composition and on the form of the solidus curve, being high in a fusible region and low in a refractory one. According our results, the temperature profile in the uppermost crust coincides with the steady-state one during, at least, 50 Ma from the initiation of the tectonic process (Fig. 11).

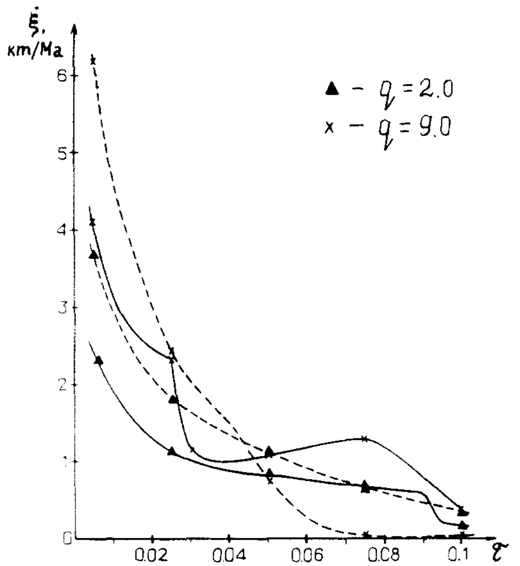

Fig. 9. Velocity of lithosphere thinning for basalt systems. Dashed lines — basalt + $H_2O + CO_2$, solid lines — basalt + H_2O, $H_2O < 0.5$ weigth %. The results for 10% of melt, $q = 2$ and $q = 9$.

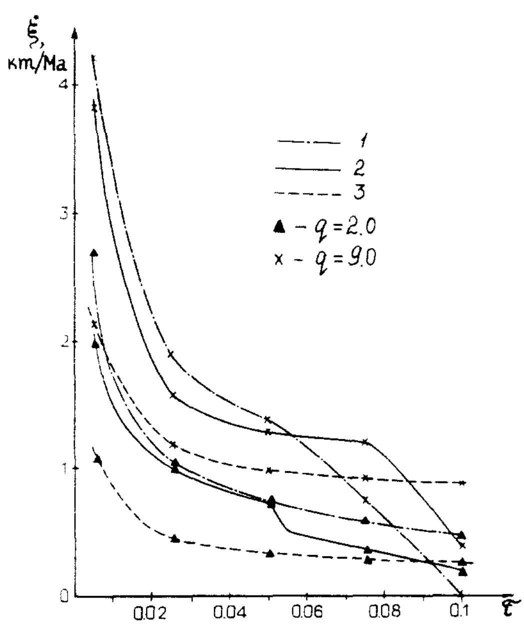

Fig. 10. Velocity of lithosphere thinning for peridotite systems: 1) peridotite + H_2O, 2) peridotite + H_2O, $H_2O < 0.5$ weigth %, 3) peridotite + CO_2. The results for 10% of melt, $q = 2$ and $q = 9$.

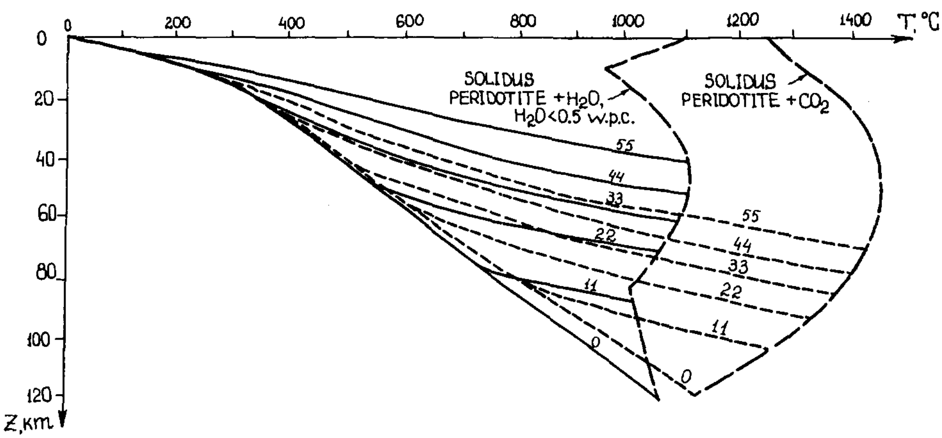

Fig. 11. Calculated geotherms for different upper-mantle compositions: solid lines — peridotite – H_2O; dashed lines — peridotite + CO_2. The results for $q = 5$, 10% of melt of lithospheric material. Numbers on the curves correspond to time (in Ma) passed from the initiation of the tectonic process; curves for O Ma show the initial steady-state temperature distribution in the lithosphere that had been used for numerical modelling.

DISCUSSION AND CONCLUSIONS

Recent geophysical investigations indicate that the upper mantle of areas of Cenozoic domal uplifts is anomalous: low-density, low-velocity, anomally heated, and partially melted. The geochemical data on incompatible elements indicate that anomalous mantle rises from a deep, undepleted reservoir. Petrological studies of Cenozoic igneous rocks have revealed a consequence in changes of magma composition with time that proves an uplift of magma sources from depths of about 100 km to the crust. Thus, we may conclude that the formation of high plateaus is induced by an ascent of mantle material to the base of the lithosphere; a small-scale convection comprising the linze of anomalous mantle may be an effective mechanism of heat transport to the base of the lithosphere. An intensive lithospheric heating leads to a decrease of the viscosity of the lithosphere material and its consecutive replacement by the anomalous mantle material.

The process of Cenozoic tectonic activation was investigated in the frame of a lithosphere-thinning model. This model is acceptable only for the areas of high lithospheric viscosity in the preactivation stage, e.g., the Tien Shan, Central Siberia mountain area, East African domal uplift. The model was described by a 1-D heat conductivity problem for the lithosphere with a moving lower boundary. The main purpose of the work was to study the influence of $H_2O - CO_2$ fluids in the upper mantle on the dynamics of lithospheric thinning.

Curves 4, 5, and 7 in Fig. 2 seem to fit present knowledge of the upper-mantle composition best. Models calculated for these cases correspond to partial melting of the roof of the upwelling diapir because of the presence of water and/or carbon dioxide. The variants 1 and 2 pertaining to different solidus conditions of the basalt system are of less interest; they were calculated mainly for comparison. Probably, these variants may be realized at the later stages of lithospheric thinning, when the lower crust becomes involved in the process.

Geophysical studies of areas of Cenozoic tectonic activity have revealed that the thickness of the crust in these regions is about 30–50 km. These areas are characterized by Cenozoic alkali-basalt volcanism generated beneath the crust. This means that a roof of the linse of anomalous mantle material (i.e., the lithosphere-anomalous mantle boundary in our model) is located deeper than 30–50 km. These data are in good correlation with numerical results (Fig. 4 and 5) according to which melting of the lower crust may occur only at the later stages of the process when the lithosphere-anomalous mantle boundary approaches its new equilibrium position and the heat flow supplied to the base of the lithosphere is large enough ($q > 3$).

If the lithosphere is essentially thinned (e.g., to the crust), the existence of deep faults may lead to intrusions of anomalous mantle material and the formations of dykes. In this case, a further development of the tectonic process can be described by a model of a "cooling dyke" (Zorin and Osokina, 1984; Logachev and Zorin, 1987), where the dyke intrusion is modelled by an instantaneous increase of temperature in a crust block of a given size and the dyke is suggested to be partially melted anomalous mantle material. Thus, we believe that the dyke model is acceptable only at the later stages of tectonic activation, while thinning of the lithosphere due to a sharp increase of a heat flow supplied to the base of the lithosphere enhances the tectonic process.

It is also important to note that ascent of anomalous mantle may lead to thickening of the crust (which is known, for example, for Tien Shan) due to two different mechanisms (Artemieva and Gliko, 1986): 1) accumulation and crystallization of basalt melts just beneath the crust, and 2) phase reactions eclogite $\rightarrow$ basalt. Phase reactions of the rocks of granulite and eclogite facias to basalt may effectively take place only under high-temperature conditions ($\sim 800°C$) (Ahrens and Schubert, 1975) which may be achieved beneath the crust only at the later stages of the thinning process. Thus, thickening of the crust may result in the acceleration of isostatic uplift of lithospheric blocks, mainly at the later stages.

Numerical modelling shows that the composition of the upper mantle strongly influences the dynamics of lithospheric thinning process. According to our results, the presence of a large amount of H_2O in the upper mantle material causes an intensive thinnig of the lithosphere and rapid isostatic uplift of lithospheric blocks, while the presence of CO_2 or a small fraction of H_2O leads to slowing-down of the thinning process.

For the calculated models of a different upper mantle composition and different values of mantle heat flow (q varying from 2 to 9), lithospheric thinning velocities lie in the interval from 1.0 to 3.0 km/Ma (5 Ma after the beginning of the tectonic process), while at the initial stage of the process a velocity of thinning may be as large as 5–10 km/Ma for models with 10% melt of anomalous mantle material. These numerical results are in a good agreement with the data on average velocities of lithospheric thinning in the areas of Cenozoic tectono-magmatic activity (2.0–2.7 km/Ma for the last 25 Ma) and data on the velocities of uplift of magma sources.

A most interesting result of our investigations is the existence of zones in the upper mantle where a sharp change of thinning velocity takes place, the velocity being increased/decreased by a factor of 2 during 1–2 Ma. Thus the depth dependence of the solidus temperature, which itself is controled by the volatile contents in the upper mantle, defines the dynamics of the process of lithospheric thermal thinning. Moreover, lamination of fusible and refractory layers in the upper mantle may lead to a discrete character of surface vertical motions which is generally caused by the decrease of the thickness of lithospheric blocks.

According to the results (Table 1), a new steady-state position of the lithosphere-anomalous mantle boundary for q varying from 3 to 5 will be reached in about 100 Ma, and 25 Ma after the initiation of the tectonic process a deflection of temperature distribution from a steady-state one exists only for depths of more than 60 km. This means that surface heat flow, as well as temperature distribution in the crust (exept zones of deep faults), may not differ from those typical for platform areas, but the usage of a steady-state heat-conductivity equation for calculation of modern geotherms in the areas of Cenozoic tectonic activation may lead to wrong results.

In modelling the process of lithospheric thermal thinning, we have assumed that the composition of the upper mantle and fluid phase is constant with depth. However, an essential vertical inhomogeneity in volatiles and in rock composition exists in real conditions of the upper mantle; this must lead to a more complicated dependence of the solidus on depth than was assumed in the model and, hence, to more a complicated character of the vertical movements of the lithospheric blocks. It should be possible in the near future to test the model for the more realistic cases.

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