ISSN 1069-3513, Izvestiya, Physics of the Solid Earth, 2007, Vol. 43, No. 1, pp. 2–12. © Pleiades Publishing, Ltd., 2007.
Original Russian Text © V.O. Mikhailov, V.M. Gordin, E.P. Timoshkina, E.A. Kiseleva, E.I. Smolyaninova, 2007, published in Fizika Zemli, 2007, No. 1, pp. 4–15.

Geodynamic Models and Their Application in the Combined Interpretation of Geological and Geophysical Data (a Review)

V. O. Mikhailov, V. M. Gordin, E. P. Timoshkina, E. A. Kiseleva, and E. I. Smolyaninova

Schmidt Institute of Physics of the Earth (IPE), Russian Academy of Sciences (RAS), Bol'shaya Gruzinskaya ul. 10, Moscow, 123995 Russia

Received June 8, 2006

Abstract—The paper presents a review of investigations in the field of the theory and practice of the interpretation of geological and geophysical data with geodynamic models that were carried out mainly by researchers of the Institute of Physics of the Earth, Russian Academy of Sciences. Evolutionary models of platform structures, passive continental margins, rift zones, and orogens are examined. The review presents formulations of inverse problems and results of interpretation for various regions, including sedimentary basins of the East European Platform, Atlantic Ocean margins, the Caucasus, the South Urals, and others.

PACS numbers: 91.45.Ga

DOI: 10.1134/S1069351307010028

INTRODUCTION

The development of this research area started with the solution of the following applied problem, which arose in 1972–1973 during the computerized construction of structural maps: how, in constructing maps of deep poorly studied structural surfaces in the sedimentary cover, should the topography of shallower well studied surfaces be taken into account? The problem was whether it is possible to find an operator linking the topography of different structural surfaces in the sedimentary cover. Veniamin P. Myasnikov posed the problem in a different way: let us first construct a geodynamic model of the formation of sedimentary cover structures, and this will allow us to determine the parameters of a tectonic process from the topography of well studied upper boundaries, after which the topography of less constrained deep boundaries can be calculated. Simultaneously, he suggested that, for the solution of the inverse problem (the determination of parameters of the model and the tectonic process), it is possible to use any available geological and geophysical data, i.e., to perform an integrated interpretation. At that time, in the early 1970s, with the low level of computer technologies, this idea seemed too adventurous. However, only a few years later, the first geodynamic models were constructed, the theory of data inversion based on geodynamic models was developed, and the first practical results were obtained. In the present paper, we present a review of the main results obtained in this field of research, which was initiated and guided for many years by Academician Myasnikov. We devote this work to the memory of this remarkable man.

FORMULATION OF THE PROBLEM

The interpretation of geophysical fields, in particular, in studies of the lithosphere and upper mantle, is usually carried out according to the following scheme. A qualitative static model of the medium under study is constructed on the basis of geological concepts and data on physical properties of rocks and is then used to determine the position and the geometry of regions differing in physical properties of the medium (density, magnetization, seismic wave velocities, etc.). In this approach, a large volume of information on the possible mechanisms of the formation and evolution of the studied structures (such as qualitative tectonic schemes, structural geology constraints on the time and patterns of tectonic deformations, data on the history of tectonic subsidence, etc.) remains unclaimed. At best, this information is invoked at a qualitative level for the formulation of interpretation models, but it is not used in the interpretation process itself. It is known that the majority of inverse problems based on separate geophysical methods belong, according to A.N. Tikhonov, to the class of conditionally correct problems. Regularization methods successfully suppress various effects of instability and equivalence, but the most reliable and natural way for achieving uniqueness and stability is the use of additional information and an integrated application of different geophysical methods.

To realize the approach based on geodynamic modeling, one should first solve the forward problem of geodynamics, i.e., construct a model describing the formation of geological structures and inhomogeneities in the distribution of physical properties that are reflected in geophysical fields [Gordin et al., 1976, 1978;

Mikhailov, 1976]. In the general case, this problem is far from its solution. Indeed, the observed variations in seismic wave velocities, density, magnetic susceptibility, and thermal conductivity are of different origins. They can be caused by tectonic movements of different scales and signs, various magmatic and postmagmatic phenomena, physicochemical transformations, and other factors. The mechanisms of the majority of these factors and, particularly, their interrelations have not been comprehensively studied, and, therefore, the construction of appropriate models in many cases is a serious problem. As regards regional tectonic structures, it is reasonable to assume that variations in physical properties are mainly due to tectonic processes displacing and deforming rocks and to thermal fields. For this reason, the thermomechanical problems could be solved within the framework of continuum mechanics. However, this required the creation of a new class of geodynamic models; these are the so-called evolutionary models, describing variations in physical properties of rocks and in the geometry of structures studied. In the early 1970s, such models were extremely scarce because, in most cases, problems were reduced to gaining analytical or numerical estimates of the distributions of velocity or stress fields. In the next sections, we consider the basic models that we used for the solution of inverse problems.

Depending on the detail of a geodynamic model and the available geological and geophysical information, various interpretation schemes are possible. We consider the following formulation of the problem, which does not claim generality. Let a geodynamic model describe the formation of one or several tectonic structures under study. This model specifies a correspondence of the initial parameters of the medium φ_i^0 with the characteristics of a tectonic process p_i, on the one hand, and the functions characterizing the present state of a structure φ_i^t and its geophysical fields and the history of their development ψ_i^t, on the other hand. The functions φ_i^0 specify initial conditions: the position of geological boundaries and the distributions of temperature, density, viscosity, elastic moduli, and so on. The velocities of tectonic motions, the heat influx from internal and external sources, and the processes of sedimentation and denudation are functions (p_i) of coordinates and time. They are mainly boundary conditions. The present structure of the medium (the functions φ_i^t) includes the topography of facial and stratigraphic boundaries; the position of faults; the distributions of seismic wave velocities, density, magnetization, and temperature; and so on. The functions ψ_i^t are seismic, gravity, and magnetic data; the heat flux; and data on the evolutionary history (tectonic subsidence rates from drilling data, fission track analysis constraints on the time and rate of exhumation of rocks, data on the degree of alteration of hydrocarbons, and so on). The functions ψ_i^t and some of the functions φ_i^t are known. Some of the initial conditions (the functions φ_i^0) can be specified a priori from general considerations or by the analogue method.

An ideal formulation of the inverse problem would be the determination of all unknown functions by minimizing the functional of misfits over all known functions with regard for the system of restraints on the functions and parameters to be determined. In practice, this scheme has not been implemented so far. Actually, the analysis of, for example, wave fields is such a time-consuming problem that, even with modern computer technologies, it is unlikely to become a part of the solution of the multidimensional minimization problem. A preferable scheme is that of successive interpretation, in which some of inverse problems are solved at the first stage, without accounting for the restraints given by the geodynamic model. For example, seismic data are used to localize seismic boundaries and determine the distribution of seismic velocities. As a result, some of the functions φ_i^t become determined. At the second stage, the following inverse problem is solved:

$$\sum_i \alpha_i \left\| \psi_{i,\,\mathrm{meas}}^t - \psi_{i,\,\mathrm{calc}}^t \right\| + \sum_i \beta_i \left\| \varphi_i^t - \varphi_{i,\,\mathrm{a\ priori}} \right\| = \min, \tag{1}$$

where the first sum includes all known data that were not used at the first stage of interpretation. The set of functions $\varphi_{i,\,\mathrm{a\ priori}}$ includes both structural characteristics of the media that were determined at the first stage and a priori information (e.g., drilling data). Equation (1) can be supplemented by restraints on the sought parameters derived from general geodynamic concepts (for example, the maximum admissible velocity of tectonic motions or the temperature) or by the conditions of proximity to given distributions (for example, one may require that densities determined from gravity data agree with the distribution of seismic velocities determined at the first stage). If a solution consistent with the available data and the geodynamic model cannot be obtained at the second stage, it is possible to return to the first stage and obtain another solution, changing the initial assumptions or interpretation parameters.

A third stage is also possible in this scheme because not all parameters of the model can be interconnected, so that some parameters are not included in the equations of geodynamics. In particular, the magnetization of rocks is not directly connected with the other characteristics. Moreover, the observed magnetic field can be produced by the distribution of magnetization in a layer of an arbitrary configuration. Therefore, magnetic data should be inverted separately, after the problem of the second stage is solved. Thus, in the integrated interpretation of geological and geophysical data for passive continental margins [Mikhailov and Trebina, 1988], the solution of problem (1) was used to update the position of the crystalline basement and the temperature distribution. Based on these data, the upper and lower boundaries of the magnetically active layer were determined and used at the third stage to solve the inversion problem of magnetometry (see below). Now, we address concrete examples of modeling and solution of inverse problems in chronological order.

Evolutionary Models of the Lithosphere and Their Application to the Solution of Inverse Problems

Initially, the approach under consideration was implemented with the use of a model of sedimentary cover deformation under the action of vertical movements of the crystalline basement surface [Zanemonets et al., 1976]. The sedimentary cover was modeled by a two-layer linearly viscous incompressible medium in the boundary layer approximation. An expansion in a small parameter made it possible to obtain a system of differential equations for the velocity field in the model layers. A distinctive feature of this model is that it converts the time evolution of the velocity field into the time evolution of the boundaries of the geological section. For this purpose, the following equation of motion of a material boundary was used:

$$\frac{\partial z}{\partial t} + U\frac{\partial z}{\partial x} = W; \tag{2}$$

i.e., it was assumed that the velocity of the boundary always coincides with the velocity of its constitutive particles. Equation (2) is written for the plane case: (x, z) are Cartesian coordinates, t is time, and U and W are the horizontal and vertical components of the velocity vector.

This geodynamic model was first used for an integrated interpretation of seismic, gravity, and drilling data for Volga–Urals structures [Gordin et al., 1976]. The problem was formulated as follows. It was assumed that the topography of the structural boundary z_1 nearest to the surface is known from seismic and drilling data. The topography of deeper boundaries of the sedimentary cover and the crystalline basement surface was to be determined from data on gravity anomalies under the assumption that there exists a deep borehole specifying the number of sedimentary layers and their asymptotes. Functional (1) included gravity anomalies alone, but it was required that the calculated upper boundary differ from z_1 by no more than a given value. It is known that the gravity inversion problem on the reconstruction of the topography of several contact surfaces does not have a unique solution. However, in this case, the topographies of all boundaries, from the basement surface to the upper stratigraphic horizon z_1, were interrelated through the equations of the geodynamic model under the assumption that the topography of sedimentary cover boundaries was produced by displacements of the crystalline basement surface. This formulation made it possible to reduce substantially the dimensionality of the inverse problem, which resulted in the uniqueness and stability of the problem. A similar formulation of the problem was used in the interpretation of seismic and gravity data for the Puchezh–Katun' meteoritic crater [Dabiza et al., 1979] in terms of a relaxation model of large impact structures in the lithosphere–asthenosphere system.

The analysis of preliminary results obtained on the basis of theoretical and practical examples showed that the accuracy of the data inversion by this approach mainly depends on how adequately the chosen geodynamic model describes the real natural process. The main factor restricting the application of the model [Zanemonets et al., 1976] is its disregard for the sedimentation and denudation processes, which are of key importance for the solution of forward and inverse problems for regional tectonic structures. A model accounting for these processes was constructed in [Mikhailov, 1983a], which significantly widened the area of possible applications. The topography of the upper boundary of the model was determined by the equation

$$\frac{\partial z_s}{\partial t} + u_s\frac{\partial z_s}{\partial x} = W_s + \varphi(x, t) + \lambda\frac{\partial^2 z_s}{\partial x^2}, \tag{3}$$

where the subscript s designates the components of the velocity vector on the surface of the model, the function $\varphi(x, t)$ specifies the rate of supply or withdrawal of material, and the last term on the right-hand side determines the redistribution of the material over the surface due to the processes of denudation and resedimentation. The type of the function $\varphi(x, t)$ is not prescribed. For the well-studied sedimentary basins, it can be specified for various time intervals according to data on the thickness of sedimentary layers. In the case of modeling passive continental margins [Mikhailov, 1983b, 1986], we assumed that the sedimentation rate exponentially decreases with the distance from the coastline. In these papers, two- and three-layer models of a linearly viscous lithosphere were examined in the boundary layer approximation. To describe the process of isostatic adjustment, a layer of an ideal fluid of a large thickness (a half-space) was placed under the lithosphere. The resulting system of equations was used to construct evolutionary models of large-scale platform depressions (for example, the Donets Basin), passive continental margins [Mikhailov, 1983b, 1986], and other regional tectonic structures. In particular, it was shown that some slow (relaxation) tectonic processes, such as the evolution of passive continental margins in the transition zone from the thicker continental crust to the thinner oceanic crust, are controlled not only by the dynamics of material motion in layers of the lithosphere and asthenosphere but also by the processes of sedimentation and denudation. With the help of new models, it was demonstrated that the isostatic compensation for a surface load can be realized through not only subsidence or uplift of the lithosphere but also variations in the thicknesses of its layers. In particular, a thickening of a sedimentary lens is accompanied by extrusion of the material of underlying crustal layers from under the sedimentary basin to its periphery. If this factor is taken into account in studies of sedimentary basins, the estimate of the lithosphere extension decreases.

Models including the processes of sedimentation and denudation have proven effective for the development of the mathematical theory of paleotectonic analysis [Myasnikov and Mikhailov, 1983; Mikhailov, 1989, 1993b], enabling the reconstruction of tectonic motion velocities from data on the thickness, age, and facial composition of sedimentary layers. First of all, the simultaneous reconstruction of the vertical and horizontal components of the velocity vector was shown to be nonunique if these functions are independent. If an equation linking these two components is known or if, for example, horizontal displacements in the region under study can be disregarded, the vertical component of the velocity can be found through the expansion in orthogonal functions of coordinates and time, in particular, in the form of a Fourier series segment. This reduces the problem to a system of linear equations, and the velocities of tectonic motions can be obtained for both sedimentation and erosion periods. The method was applied to the analysis of the formation of the Terek-Caspian trough [Mikhailov, 1993a] and the Moscow basin [Mikhailov et al., 2006].

A large volume of new information, in particular, data on the lithologic–facial composition of sedimentary rocks, could be incorporated into the interpretation of geophysical data in terms of the new models. Now, we address an example of combined analysis of seismic, gravity, and magnetic data, as well as data on tectonic subsidence rates at passive continental margins [Mikhailov and Trebina, 1988]. The solution scheme was the following. In the geodynamic model used for interpretation, the structure and evolution of the continental–oceanic lithosphere transition zone is determined by a set of parameters including the characteristic vertical and horizontal scales of the structure studied; the densities and the effective viscosities of layers of the sedimentary cover, crust, and upper mantle; and the process duration (the age of the margin). Functional (1) included the norm of the misfit between the calculated topography of sedimentary upper layers and the available seismic data, as well as the norm of the deviation of the calculated tectonic subsidence rate from its actual value known from drilling data. It was also necessary to adjust the amplitude and the characteristic dimension of the calculated isostatic anomaly to their observed values. The problem was solved along a profile crossing the eastern continental margin of North America. As a result, the contemporary configuration of all layers was calculated, including the basement and Moho discontinuity surfaces, which are poorly known from seismic data due to the large thickness of sediments and the presence of strong reflectors in them. The data on the crystalline basement surface and the temperature distribution were then used for the localization of the upper and lower boundaries of the magnetically active layer, which made it possible to estimate the distribution of magnetization from magnetic anomalies with the help of the method described in [Gordin et al., 1987]. Based on the comparison of the inferred magnetization distribution with the position of extension and compression zones, the formation of the Brunswick magnetic anomaly at the eastern margin of the United States was associated with the emplacement of intrusive material into an extension zone on the inner (landward) flank of the sedimentary basin after the rifting stage. Note that basalts of two generations dated at 138–112 and 75–40 Ma were discovered in drillcores at the eastern continental margin of South America, which has a similar structure. This confirms the conclusion formulated on the basis of modeling results according to which, although the major extension of the lithosphere occurs at the rifting stage, the formation of a sedimentary basin at the postrifting stage is accompanied by additional extension significantly increasing the depth of the basin.

The role of the sedimentation process was also studied using as examples the Gakkel Ridge [Mikhailov and Timoshkina, 1993] and the subsidence history of the Great Valley (California) lithosphere in the delta of the Sacramento River [Mikhailov et al., 2007]. In the latter case, the thermal regime and the subsidence history of the forearc basin on the ~150-Ma oceanic lithosphere was modeled. Here, it was necessary to model not only the cooling of the oceanic lithosphere overlain by a thick (up to 14–16 km) sedimentary cover but also the thermal regime of the subduction zone, taking into account time variations in the subduction rate and the slab age. To describe the cooling and subduction of the oceanic lithosphere, we used the model [Mikhailov and Timoshkina, 1993] in which the crystallization heat of a basaltic melt at the lithosphere–asthenosphere boundary and a sedimentary layer thickening with time are taken into consideration. The constructed thermal model is in good agreement with data on the present-day heat flux, the history of tectonic subsidence, and the thermal alterations of organic matter of sedimentary rocks. The inferred contemporary nonstationary temperature distribution in the lithosphere was used for the calculation of an yield strength profile. The strength profile predicts the presence of a brittle layer in the upper crust that can extend to a depth of 20 km or more. This explains the presence of an anomalously deep (up to 20 km) cluster of earthquakes in the area of the Sacramento Delta. The depth of earthquakes does not exceed 12 km in adjacent areas that differ in crustal structure, sedimentary cover thickness, and heat flux. Using the considered approach for joint analysis of seismic, drilling, and thermal data, as well as results of plate tectonic reconstructions, the paper cited above represents, in our opinion, an interesting example of the application of a geodynamic modeling technique and basin analysis to seismological problems.

Another example of application of this approach is the analysis of global time variations in the gravity field based on data of the GRACE satellites. As was shown in [Mikhailov et al., 2004], temporal variations in the gravity field related to earthquakes with magnitudes greater than 9 (e.g., Chile, 1960, and Alaska, 1964) are comparable with the uncertainties of gravity models derived from GRACE data and, therefore, are not always visually detectable. On the other hand, ground-based data can generally be used to construct a model of the displacement in an earthquake focus, for example, with the help of the dislocation model in an elastic half-space. The rupture zone is approximated by a set of planes whose size, position, and slip amplitudes are determined from geodetic data on the displacement of the Earth's surface. The following two problems were examined in [Mikhailov et al., 2004]: (i) detection of a gravity signal from an earthquake, provided that the signal shape is known up to a constant (i.e., the evaluation of the displacement in a source whose sizes and position are known), and (ii) recognition of a satellite-derived gravitational signal consistent with seismic source models equivalent in their fit of seismological and geodetic data. The statistical recognition procedure proposed in the paper cited above is effective for the solution of both problems. Diament et al. [2007] applied this theoretical approach to the analysis of the temporal variations in the gravity field due to the Andaman–Sumatra earthquake of December 2004. It was shown that, together with a gravitational signal from the displacement in the rupture zone, there exists a negative anomaly in the southern Andaman Sea that can be related to a change in the density of rocks in the lithosphere and mantle as a result of deformation (dilatancy) or to subsidence of this region as a result of the earthquake. The estimated amplitude of subsidence is ±15 cm. The above problems could be solved due to the integrated approach to the interpretation of ground-based and satellite data within the framework of geodynamic models.

An Evolutionary Model of a Rheologically Stratified Outer Shell of the Earth

New possibilities in the modeling of geodynamic processes and the interpretation of regional geophysical data arose after the development of a thermomechanical evolutionary model of a rheologically stratified outer shell of the Earth adjusted to the hydrodynamic model of the Earth's evolution. The model includes the lithosphere (consisting of the sedimentary cover, crust, and subcrustal mantle), the asthenosphere, and a part of the upper mantle below the asthenosphere. We consider briefly the basic principles of the construction of this model; its details are presented in [Myasnikov et al., 1993; Mikhailov et al., 1996a; Timoshkina, 1998].

To describe motions in the bulk mantle volume, three characteristic parameters were introduced [Myasnikov and Fadeev, 1980]: the Rayleigh number $\mathrm{Ra} = g_0R^3\rho_0/(\kappa_0\eta_0)$, the time of convective mixing of material $t_0 = R^2/(\mathrm{Ra}^{1/2}\kappa_0)$, and the heat transfer rate $v_0 = R/t_0$. Here, g_0 is the gravity acceleration; R is the Earth's radius; and ρ_0, η_0, and κ_0 are the mean density, viscosity and thermal diffusivity of the mantle, respectively. The following values were accepted for the main volume of the mantle: $\rho_0 = 3 \times 10^3$ kg/m^3, $\eta_0 = 10^{23}$ Pa s, and $\kappa_0 = 10^{-6}$ m^2/s; hence, given $R = 6.4 \times 10^6$ m and $g_0 = 10$ m/s^2, we obtain $\mathrm{Ra} = 10^8$. This allows us to introduce the small parameter $\varepsilon = 1/\sqrt{\mathrm{Ra}} \approx 10^{-4}$ and to obtain a determining system of equations in the main volume of the mantle after the introduction of dimensionless variables and the successive application of the expansion in the small parameter to the Stokes equations.

The outer shell of the planet is considered as a thermal boundary layer consisting of four layers of constant viscosities:

boundary layer of the upper mantle with the viscosity $\eta_m = \eta$;

low viscosity asthenosphere, $\eta_a = \sqrt{\varepsilon}\,\eta$;

high viscosity lithosphere, $\eta_1 = \eta/\sqrt{\varepsilon}$;

low viscosity sedimentary cover, $\eta_s = \sqrt{\varepsilon}\,\eta$.

To obtain equations in the boundary layer in the neighborhood of the mantle surface, a stretched vertical coordinate (Z) is introduced. The origin of coordinates in the boundary layer is placed under the asthenosphere base: $Z - R_0 = (z - R)/\sqrt{\varepsilon}$, where R_0 is the position of the hydrodynamic radius in the boundary layer (stretched) coordinate system. The concept of the hydrodynamic radius is analogous to the concepts of the floating level or the free mantle level, used in geodynamics. The system of equations for the boundary layer is obtained from the equations for the mantle after the introduction of the stretched coordinate (the vertical velocity component in this case is, accordingly, $W = w/\sqrt{\varepsilon}$) expanded in powers of $\sqrt{\varepsilon} \cong 10^{-2}$ and the introduction of boundary conditions at the outer and inner boundaries.

The adjustment of the solution in the boundary layer to the solutions in the main volume of the mantle is of basic importance. To do this, the method of splicing of asymptotic expansions was applied [Myasnikov and Savushkin, 1978]. According to this method, the following relation of asymptotic equivalence should be valid near the model surface ($z = R$) for any function f represented as an expansion in $\sqrt{\varepsilon}$ in the boundary layer ($f^{(i)}$) and as an expansion in ε in the main volume of the model (f'):

$$[f^{(0)} + \sqrt{\varepsilon} f^{(1)} + \dots]_{z \to -\infty} \cong \left[f' + \sqrt{\varepsilon}(Z - R_0)\frac{\partial f'}{\partial z} + \dots \right]_{\varepsilon \to 0,\, z = R}, \quad (4)$$

where R_0 is the hydrodynamic radius of the Earth in the stretched coordinate system.

The above conditions of the adjustment to the global model yielded the conditions that must be satisfied in the boundary layer and, in particular, the equation linking the vertical and horizontal velocities at the base of the boundary layer and the condition of global isostasy.

In the boundary layer model constructed, the layers differ in density and effective viscosity, the sedimentation and denudation processes are included in the evolutionary equation of the Earth's surface, and the top and the base of the asthenosphere can be either material or rheological boundaries (in the latter case, they are defined by a certain isotherm, typically yielding 1300–1350° for the asthenosphere top).

The qualitative analysis of the inferred equations and numerical calculations [Timoshkina, 1998] suggest that violations of the mechanical and thermal equilibrium in the outer shell lead to the formation of circulating flows in the asthenosphere that persist for a long time after the termination of the active stage, i.e., the period of action of external tectonic forces. Based on this model, the development of small-scale convection could be examined for the first time in extension and compression zones formed due to intraplate or mantle effects.

To study the interaction between the lithosphere and asthenosphere, we used the widely accepted scheme of the development of tectonic structures, which includes the following two stages. At the initial stage of active tectogenesis a few million years long, intense deformation of the outer shell occurs (for example, extension or compression by horizontal intraplate forces or by the forces due to activation of the underlying upper mantle). Active tectonic processes lead to violation of the mechanical and thermal equilibrium in the lithosphere–asthenosphere system, which starts recovering at the second, longer stage, mainly after the termination of the action of external tectonic forces. It is important that, although the viscosity of the lithosphere is four orders of magnitude higher than that of the asthenosphere, convective flows in the asthenosphere have a significant effect on the lithosphere: extension zones form above ascending asthenospheric flows, and compression zones form above descending ones. As a result, subsidence and uplift regions arise that, in combination with the sedimentation and denudation processes, significantly affect the structure and evolution of tectonic units.

In the absence of external tectonic forces (the relaxation stage), the evolution of tectonic structures is determined by several factors, including the density and temperature distributions in the outer shell, the width of the extension or compression zones, and the intensity of sedimentation and denudation. It is important that, if the density in the asthenosphere does not decrease with depth, small-scale flows maintain mainly the same deformation pattern as at the active stage: small extension and low amplitude subsidence continue in extension zones, while compression and uplift persist in compression zones [Timoshkina, 1998]. These effects should be taken into account in the analysis of the formation history of sedimentary basins and mountain belts. Below, we present some new results obtained within the framework of this model.

The relationship between the value of the initial extension of the lithosphere and the thickness of a sedimentary basin is determined by the sizes of the extension region and by the intensity of small-scale convection in the asthenosphere. Although many works have been devoted to the modeling of the lithosphere extension process, many aspects of this problem have not been adequately studied as yet. In particular, the sedimentation and denudation effects, as well as processes related to the recovery of the mechanical equilibrium in the lithosphere–asthenosphere system, have not been analyzed in detail.

Now, we address the next example. We assume that, at the initial time, the layers of the Earth's outer shell were horizontal and their thicknesses amounted to 0 km for the sedimentary cover (older deposits are included in the lithosphere composition), 100 km for the lithosphere (including the crustal layer 36 km thick), and 100 km for the asthenospheric layer. At the initial time moment, the density within each layer depended only on depth. The density of the newly forming sedimentary layers was assumed to be constant (2.4 g/cm^3) and did not vary in time. The density of the crust increased with depth from 2.7 g/cm^3 at the top to 2.9 g/cm^3 at the base, the density of the subcrustal lithosphere was set equal to 3.35 g/cm^3, and the density of the asthenosphere increased from 3.35 g/cm^3 at its top to 3.36 g/cm^3 at the base. The initial temperature distribution was assumed to be stationary.

At the first (active) stage, 4 Myr long, the rate of extension by external forces was specified in a way ensuring uniform extension in a region 480 km wide (the process modeling the formation of a backarc basin). The maximum rate at the boundaries of the extension region reached 7 mm/yr, and the overall extension value over 4 Myr was 56 km, or about 12%. In the extension period, the crustal thickness decreased to 32 km and the depth of the resulting sea basin was 800 m.

At the first stage, during the extension, the thicknesses of the outer shell layers and the temperature distribution vary in such a way that the isotherms under the extension region rise and the temperature distribution becomes nonstationary. As a result, horizontal gradi-

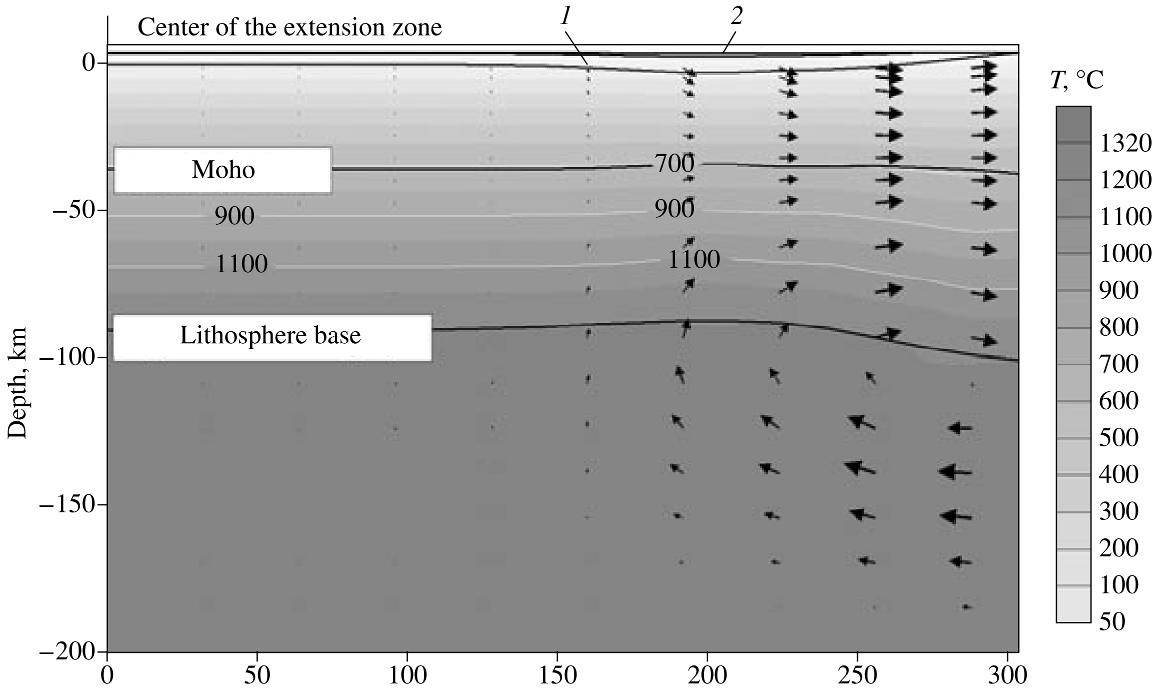

Fig. 1. Small-scale convection in the asthenosphere 80 Myr after the termination of the phase of extension by intraplate forces. The arrows indicate the direction of motion. The maximum arrow size corresponds to a velocity of 0.18 mm/yr. The white contours and shades of gray show the temperature distribution: (*1*) sedimentary layer formed in the first 60 Myr after the termination of extension; (*2*) sedimentary layer formed within the interval 60–80 Myr after the termination of extension.

ents of the lithostatic pressure arise in the lithosphere and asthenosphere due to both variations in the temperature of the layers (the model accounts for the temperature dependence of the rock density) and the rock displacement (since the rock density increases with depth, the uplift or subsidence of layers in the process of extension or compression disturbs the initial mechanical equilibrium). As the disturbances increase, small-scale convective flows persisting for a long time develop in the low viscosity asthenosphere. The motions are transmitted from the asthenosphere into the lithosphere.

As noted above, the velocity and direction of convective motions significantly depend on the initial density distribution over depth [Timoshkina, 1998]. In particular, in the example given in Fig. 1, the asthenosphere density increases with depth by 0.01 g/cm^3. With this density distribution, the asthenosphere material at the relaxation stage rises under an extension region and subsides on its periphery. This causes additional extension and subsidence in the sedimentary basin region and small compression (uplift) near its boundaries. Convective motions continue to affect the lithosphere for a long time after the application of external forces, although the motions become much slower: in the given example, the maximum velocity of the motions at this stage is 0.2 mm/yr.

An important result of this study is the fact that a considerable depth of the sedimentary basin (about 10 km) is obtained for a relatively small initial coefficient of expansion. This fact is of basic importance because, in the widespread model of McKenzie [1978], the observed thickness of a sedimentary basin is often obtained with extension values that are appreciably larger than the estimates derived from variations in the crustal thickness or from reconstructions of systems of listric faults (e.g., see [Ziegler, 1992]). This is due to the fact that, in McKenzie's model, extension occurs at the initial (active) stage alone. At this stage, the velocities of tectonic motions are high and the so-called synrifting sediments are deposited in narrow grabens. The subsequent stage of slow thermal subsidence is due to cooling of the lithosphere, whose temperature rose, due to the extension, above the stationary value. At this stage, the so-called postrifting sediments are deposited, and their thickness in many cases is smaller than the estimate obtained in this model from the observed value of extension. Our model demonstrates that the extension is partly produced by small-scale convection in the asthenosphere developing after the termination of the active stage. As a result, the required initial extension is considerably smaller and the synrifting deposits become thinner, but this leads to an increase in the thickness of the postrifting deposits. Thus, in the above example, the extension at the active stage was 12% and, 100 Myr after the termination of the active stage, nearly doubled in the deepest part of the basin. Since the related strain rates were small, the additional extension could mainly occur in the creep mode, without clearly expressed faulting. (Here we should note that many seismic profiles and outcrops fix ruptures that formed much later than the termination of the active extension stage.)

Figure 2 plots the subsidence curves for model points at which, due to the small-scale convection, the extension increased over 100 Myr to 21% (curve *2*) and 24% (curve *1*). The subsidence curve constructed from the model of McKenzie [1978] with an initial extension value of 30% (curve *3*) and two subsidence curves derived from drilling data of boreholes in the North Sea [Sclater and Christie, 1980] (curves *4*, *5*) are also presented in Fig. 2. It is evident that the subsidence curves calculated with regard for small-scale convection give larger subsidence with smaller extension and agree better with the data on the subsidence process of the North Sea basin.

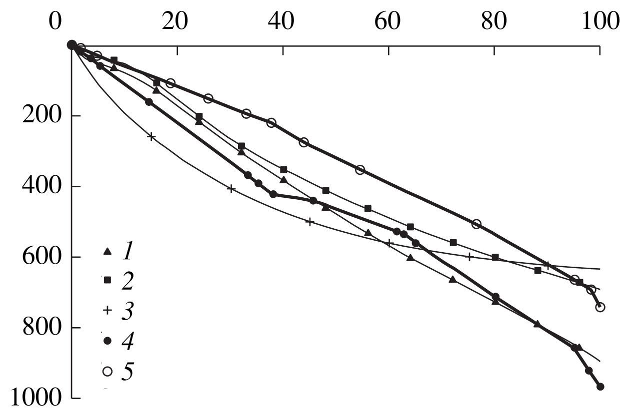

Fig. 2. Subsidence curves for a sedimentary basin obtained with regard for small-scale convection in the asthenosphere: (*1*, *2*) theoretical subsidence curves at points where the total extension value amounted to 21 (*1*) and 24% (*2*); (*3*) model of McKenzie [1978], with an extension value of 30%; (*4*, *5*) the BP 30/1-1 (*4*) and Montrose Amoco 22/18-2 (*5*) subsidence curves for the North Sea [Sclater and Christie, 1980]. The subsidence depth in meters is plotted on the vertical axis, and the time (in Myr) is plotted on the horizontal axis.

Formation of deep sedimentary basins in the continent–ocean transition zone. The horizontal dimension of a convective cell is close to the asthenosphere thickness. Therefore, if the trough is sufficiently wide, convection in the asthenosphere involves not the entire region of extension but only its periphery, i.e., the transition zone from the stretched lithosphere of the newly formed sedimentary basin to the normal continental lithosphere (see Fig. 1). As a result, the region of maximum subsidence of the sedimentary basin shifts toward its periphery, forming there a deeper trough. The structure of the upper part of the sedimentary cover is shown in Fig. 3. As in the preceding example, the subsidence is significantly larger than the thermal subsidence in the model of McKenzie [1978], which makes the depth of sedimentary basins at passive continental margins (e.g., the western Atlantic margin) consistent with the observed value of extension of the lithosphere underlying these basins.

Formation of troughs on the periphery of compressional orogens. Calculations show that the violation of mechanical and thermal equilibrium in compression regions of the continental lithosphere by the intraplate forces leads to the formation of small-scale flows in the asthenosphere beneath peripheral zones of the compression regions [Mikhailov et al., 1999b]. These flows produce additional compression in the orogen and extension on its periphery, contributing to the formation of foredeeps. Subsidence on the periphery of compressional orogens is often related to elastic bending of the lithosphere under the weight of a mountain structure, but, in many cases, the topography effect is insufficient for the creation of the observed trough. Therefore, for many mountain structures, the presence of a so-called hidden load, a positive density anomaly within the crust, is assumed. For a number of mountain structures (e.g., the Greater Caucasus), this assumption is unacceptable because gravity anomalies are inconsistent with the presence of higher density inclusions in the crust.

We have applied the evolutionary model of the outer shell to an integrated analysis of data for the Northern Caucasus [Mikhailov et al., 1999b]. The thicknesses of outer shell layers, the distribution of physical properties and temperature, the width of the compression region, the amplitude of compression, and other parameters of the model were chosen through their adjustment to the present-day data on the structure of the mountains and foredeeps and to the tectonic subsidence rates estimated from logs of more than 100 boreholes [Mikhailov et al., 1999a]. It is important that, in the paper cited, the comparison of the main events in the history of subsidence of Caucasus foredeeps with the phases of tectonic compression and volcanism in the North Caucasus region showed that uplift phases in a foredeep corresponded to compression phases in the orogen, and the subsidence periods, to noncompressional phases. Precisely this pattern is obtained from the evolutionary model of the outer shell, whereas the model of elastic bending of the lithosphere predicts an opposite correlation: the phase of external tectonic compression is accompanied by thrusting of the orogen over the foredeep region and, as a consequence, by the subsidence of the latter.

Application of geodynamic modeling to the numerical estimation of crustal stresses. We restrict ourselves to the 2-D variant of the problem. For example, let a model of deep structure and the depth distribution of physical properties be obtained from integrated analysis of geological and geophysical data gathered on a certain seismic profile. Using these data, stresses can be numerically estimated under certain boundary condi-

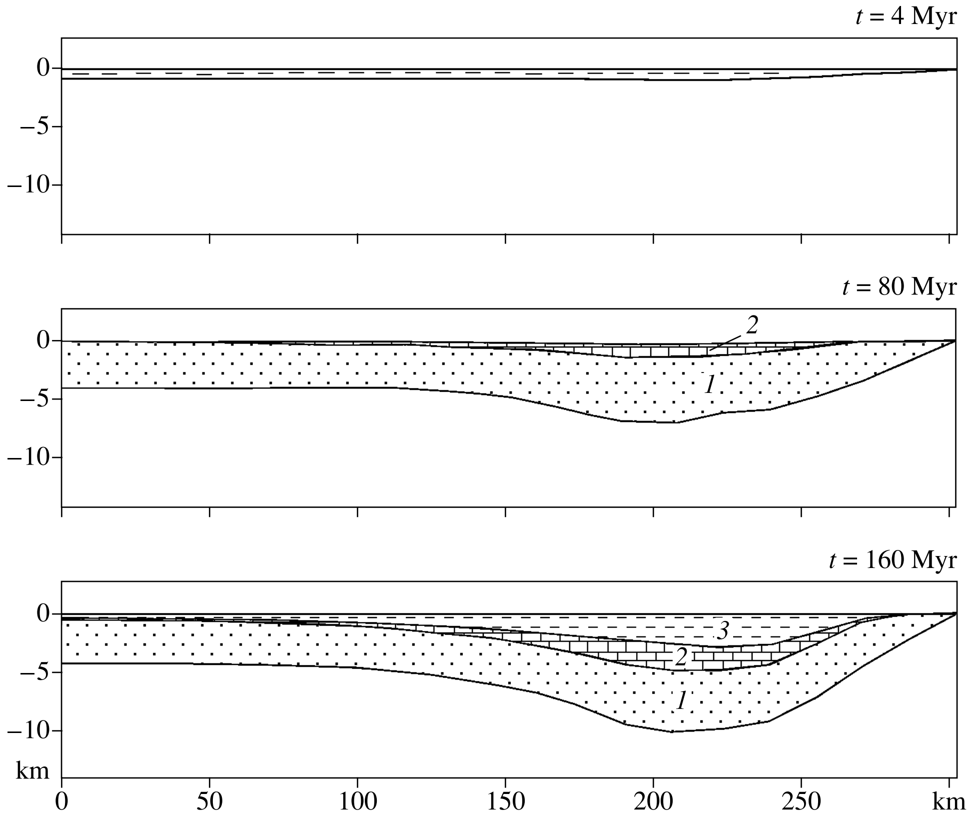

Fig. 3. Structure of sedimentary layers on the periphery of a wide sedimentary basin for the example shown in Fig. 1. The numbers indicate sedimentary layers formed in the first 60 Myr after the termination of extension by intraplate forces (layer *1*) and in the intervals 60–100 Myr (layer *2*) and 100–160 Myr (layer *3*).

tions, i.e., tectonic forces applied to the lateral boundaries (intraplate forces) and the base of the model (mantle forces). However, the existing qualitative ideas on the type of the acting tectonic forces preclude the specification of numerical values for boundary conditions. On the other hand, data on recent vertical and horizontal crustal movements, including geodetic (GPS) and neotectonic evidence, remain unclaimed. These data can be used for solving the following inverse problem: the boundary conditions in the chosen structural model of a certain region of the lithosphere (defined at its lateral boundaries and the base) are set in such a way that the velocities of surface movements are close in their norm in the chosen metric to velocities known from geodetic and/or neotectonic constraints. If the boundary conditions are represented as an expansion over a system of basic functions, the inverse problem reduces to the determination of the corresponding expansion coefficients. This approach was proposed and realized for the first time for a profile crossing the Kola Peninsula [Kolpakov et al., 1991]. At present, this approach has been applied to the estimation of stresses in a number of regions on the basis of data from profiles crossing the Black Sea and the Crimea, the Beaufort Sea, the Greater Caucasus and the North Caucasus region (the region of the Azov-Kuban and the Terek-Caspian troughs and the Stavropol uplift), the Lesser Caucasus, and the South Urals (see the bibliography in [Smolyaninova et al., 1996, 1997; Mikhailov et al., 2002a, 2002b]).

In particular, the crustal structure for the South Urals was specified from the data of the Urseis profile, and the density distribution, from the results of interpretation of gravity anomalies. The distribution of mechanical parameters was determined on the basis of yield strength diagrams calculated from data on the composition of rocks and the temperature distribution in the crust. In order to set boundary conditions, we used the amplitude of vertical movements at the neotectonic stage. As a result, it was shown [Mikhailov et al., 2002b] that the present-day topography of the South Urals could have formed due to simple intraplate compression, while the strain distribution was determined by the heterogeneous crustal structure and temperature distribution. It is important to note that the numerical model predicts decoupling of the rigid layers in the upper crust and the lower crust layers, separated by the layer of the middle crust, where creep deformation occurs. Maximum strains in the upper crust concentrate in the Main Ural fault zone; in the lower crust, they are displaced by 70 km to the west and are fixed beneath the Western Ural uplift. It is precisely in this region that a jump in the Moho discontinuity was discovered (the so-called Makarovskii fault), which is not traceable toward the surface and has no signatures in the structure of the upper part of the geological section [Mikhailov et al., 2001].

CONCLUSIONS

This paper presented a review of studies performed mainly by IPE RAS researchers in the field of the theory and practice of geological and geophysical data interpretation within the framework of geodynamic models. In our opinion, the results obtained thus far have revealed a great potential of this research area, whose development is inseparably linked with Veniamin P. Myasnikov.

REFERENCES

1. A. I. Dabiza, M. S. Krass, V. O. Mikhailov, et al., "Geophysical and Geomechanical Aspects of the Study of Meteorite Structures," Phys. Earth Planet. Inter. **20**, 1–11 (1979).
2. M. Diament, G. C. P. King, V. Mikhailov, et al., "Satellite Gravity Reveals Subsidence in the Andaman Sea Caused by Giant Sumatra Earthquake," Geophys. Res. Lett. (2006).
3. V. M. Gordin, V. O. Mikhailov, and V. P. Myasnikov, "Perspectives of the Interpretation of Gravity Anomalies Using Models of Structural Formation," in *Proc. IV Moscow Seminar on the Theory and Practice of Interpretation of Gravity and Magnetic Anomalies* (1976), pp. 1–15.
4. V. M. Gordin, V. B. Zanemonets, V. O. Mikhailov, and V. P. Myasnikov, "Mechanical Modeling of the Formation of Crustal Structures in Geological Interpretation of Regional Gravity Anomalies," Geofiz. Sb., No. 81, 22–39 (1978).
5. V. M. Gordin, V. O. Mikhailov, and E. S. Trebina, "Methods and Results of Interpreting Magnetic Anomalies in South Atlantic," Izv. Akad. Nauk SSSR, Ser. Fiz. Zemli, No. 7, 69–83 (1987).
6. I. T. Kolpakov, V. A. Lyakhovsky, M. V. Mints, et al., "Geodynamic Origin of Relief-Forming Processes on the Kola Peninsula," Geotektonika **25**, 161–166 (1991).
7. D. P. McKenzie, "Some Remarks on the Development of Sedimentary Basins," Earth Planet. Sci. Lett. **40**, 25–31 (1978).
8. V. O. Mikhailov, "Application of Mechanical Models of Structural Formation to Gravity Data Inversion," Region. Razved. Promysl. Geofiz., No. 15, 1–17 (1976).
9. V. O. Mikhailov, "A Mathematical Model of the Evolution of Structures Forming Due to Vertical Movements," Izv. Akad. Nauk SSSR, Fiz. Zemli, No. 6, 3–18 (1983a).
10. V. O. Mikhailov, "A Dynamic Model of the Evolution of Passive Continental Margins," in *Problems of Mechanics of Natural Processes* (MGU, Moscow, 1983b), pp. 155–175 [in Russian].
11. V. O. Mikhailov, "A Mathematical Model of the Formation of Sedimentary Basins at Continental Margins of the Atlantic Type," in *Structure and Dynamics of Continent–Ocean Transition Zones* (Nauka, Moscow, 1986), pp. 89–96 [in Russian].
12. V. O. Mikhailov, "A Mathematical Method of Solving the Problem of Paelotectonic Analysis," Izv. Akad. Nauk SSSR, Fiz. Zemli, No. 3, 78–90 (1989).
13. V. O. Mikhailov, "Crustal Control of the Terek-Caspian Trough Evolution: Constraints Based on a New Paleotectonic Analysis Method," Tectonophysics **228**, 21–32 (1993a).
14. V. O. Mikhailov, "A Method to Solve Paleotectonic Analysis Problem," Math. Geol. **25** (7), 949–961 (1993b).
15. V. O. Mikhailov and E. S. Trebina, "Geophysical Fields and a Model of the Evolution of Passive Continental Margins," Izv. Akad. Nauk USSR, Ser. Geofiz. **10** (5), 12–19 (1988).
16. V. O. Mikhailov and E. P. Timoshkina, "Analysis of Data on the Gakkel Ridge on the Basis of a Thermal Model of an Oceanic Rift Zone," Dokl. Akad. Nauk **331**, 497–499 (1993).
17. V. O. Mikhailov, V. P. Myasnikov, and E. P. Timoshkina, "Dynamic Evolution of the Earth's Outer Shell under Extension and Compression," Izv. Akad. Nauk, Fiz. Zemli, No. 6, 30–37 (1996a) [Izvestiya, Phys. Solid Earth **32**, 44496–502 (1996a)].
18. V. O. Mikhailov, E. P. Timoshkina, and E. A. Kiseleva, "Analysis of the Evolution of Sedimentary Basins Using a Model of a Rheologically Stratified Surface Shell," in *Problems of Continuum Mechanics*, Ed. by V. A. Levin and M. A. Guzev (Vladivostok, 1996b), pp. 142–159 [in Russian].
19. V. O. Mikhailov, L. V. Panina, R. Polino, et al., "Evolution of the North Caucasus Foredeep : Constrains Based on the Subsidence Curves Analysis," Tectonophysics **308**, 361–380 (1999a).
20. V. O. Mikhailov, E. P. Timoshkina, and R. Polino, "Foredeep Basins: The Main Features and Model of Formation," Tectonophysics **308**, 345–360 (1999b).
21. V. O. Mikhailov, E. A. Kiseleva, E. I. Smolyaninova, et al., "Estimation of Regional and Local Stresses along the Uralseis Profile," in *Deep Structure and Geodynamics of the South Urals (the Uralseis Project)*, Ed. by A. F. Morozov (Tver, 2001), pp. 275–283 [in Russian].
22. V. O. Mikhailov, E. I. Smolyaninova, and M. Sebrier, "Numerical Modelling of Neotectonic Movements and State of Stress in the North Caucasus Foredeep," Tectonics **21**, 1029 (2002a).
23. V. O. Mikhailov, A. V. Tevelev, A. G. Berzin, et al., "Constraints on the Neogene–Quaternary Geodynamics of the Southern Urals: Comparative Study of Neotectonic Data and Results of Strength and Strain Modeling along the URSEIS Profile," in *Mountain Building in the Uralides: Pangea to Present* (Geophysical Monograph 132, 2002b, pp. 273–286).
24. V. Mikhailov, S. Tikhotsky, M. Diament, et al., "Can Tectonic Processes Be Recovered from New Gravity Satellite Data?," Earth Planet. Sci. Lett. **228** (3–4), 281–297 (2004).
25. V. O. Mikhailov, E. A. Kiseleva, E. P. Timoshkina, et al., "Numerical Geodynamic Analysis of the Formation History of the Moscow Basin in the 1EB Profile Cross Section: A Formation Scenario and the Assessment of Oil

and Gas Content," in *Methodological Aspects of Research on the Internal Structure and Mineral Resources of the East European–Barents Platform Megaprovince* (MPR RF, Moscow, 2006) [in Russian].

26. V. O. Mikhailov, T. Parsons, R. Simpson, et al., "Why the Sacramento Delta Area Differs from Other Parts of the Great Valley: Numerical Modeling of Thermal Structure and Thermal Subsidence of Forearc Basins," Fiz. Zemli, No. 1, 80–95 (2007) [Izvestiya, Phys. Solid Earth **43**, 75–90 (2007)].
27. V. P. Myasnikov and V. D. Savushkin, "Method of Small Parameter in a Hydrodynamic Model of the Earth's Evolution," Dokl. Akad. Nauk SSSR **238**, 1083–1086 (1978).
28. V. P. Myasnikov and V. E. Fadeev, *An Evolutionary Model of the Earth and Terrestrial Planets* (VINITI Ser. Itogi Nauki, Moscow, 1980), Vol. 5 [in Russian].
29. V. P. Myasnikov and V. O. Mikhailov, "On the Mathematical Theory of Paleotectonic Analysis," Dokl. Akad. Nauk SSSR **270** (4), 851–855 (1983).
30. V. P. Myasnikov, V. O. Mikhailov, and E. P. Timoshkina, "On the Coupling of the Mantle with a Rheologically Stratified Outer Shell of the Earth," Dokl. Akad. Nauk **330**, 771–773 (1993).
31. J. G. Sclater and P. A. F. Christie, "Continental Stretching: An Explanation of the Post-Mid-Cretaceous Subsidence of the Central North Sea Basin," J. Geophys. Res. **85** (B7), 3711–3739 (1980).
32. E. I. Smolyaninova, V. O. Mikhailov, and V. A. Lyakhovsky, "Numerical Modelling of Regional Neotectonic Movements in the Northern Black Sea," Tectonophysics **266**, 221–231 (1996).
33. E. I. Smolyaninova, V. O. Mikhailov, and V. A. Lyakhovsky, "Numerical Modeling of Regional and Local Fields of Stresses in the Northern Part of the Black Sea," Izvestiya, Phys. Solid Earth **33**, 322–328 (1997).
34. E. P. Timoshkina, A Mathematical Evolutionary Model of an Earth's Rheologically Stratified Outer Shell, *Extended Abstract of Cand. Sci. (Phys.–Math.) Dissertation*, OIFZ RAN, Moscow, 1998, p. 25 [in Russian].
35. V. B. Zanemonets, V. O. Mikhailov, and V. P. Myasnikov, "A Mechanical Model of Block Folding Formation," Izv. Akad. Nauk SSSR, Ser. Fiz. Zemli, No. 10, 13–23 (1976).
36. P. Ziegler, "Europe Cenozoic Rift Zones," Tectonophysics **208**, 91–111 (1992).