Evolution of mantle plumes and uplift of continents during the Pangea breakup

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[1] The paper presents results derived from numerical modeling of mantle heating and reorganization of mantle flows during assemblage of two continents and subsequent breakup of the supercontinent. The simplest mantle model consisting of an extended rectangular region filled with a viscous fluid heated from below is considered. Thermal convection develops in the mantle. Due to viscosity dependence on temperature and pressure, a high viscosity lithosphere and a low viscosity asthenosphere arise in the mantle. Two continents modeled as rigid thick floating plates are then placed into the mantle. Driven by viscous coupling with mantle flows, the continents start moving and converge above the nearest descending mantle flow. The resulting supercontinent hinders the outflow of mantle heat, the subcontinental mantle starts heating, and the cold descending mantle flow gives way to a hot ascending flow. The latter assumes the shape of a plume with a spherical head and a thin tail. The resulting tensile stress breaks up the supercontinent. The Atlantic Ocean structure with a ridge and a lithosphere thickening toward continents forms between the diverging parts of the supercontinent. The calculated distribution of heat flux from the mantle has a maximum of about 200 mW m^{-2} in the zone of the mid-ocean ridge and drops by about six times toward the continents, which agrees with data of observations. The Pacific Ocean structure with typical subduction zones develops on the opposite side of the continents. A high-viscosity mantle layer corresponding to continental lithosphere moves together with continents. The Atlantic Ocean structure persists for about 50–100 Myr, after which the heavy oceanic lithosphere separates from the continents and starts sinking into the mantle. Both continents are uplifted by a few hundred meters for about 50 Myr before and 50 Myr after the breakup of the supercontinent. This result is consistent with data on sea level fluctuations and can account for the duration of the Paleozoic, Mesozoic, and Cenozoic periods and for the origin of their changes.   *INDEX TERMS:* 8120 Tectonophysics: Dynamics of lithosphere and mantle: general; 8121 Tectonophysics: Dynamics: convection currents, and mantle plumes; 8149 Tectonophysics: Planetary tectonics; 8158 Tectonophysics: Plate motions: present and recent; *KEYWORDS:* Global geodynamics, Numerical modeling, Convection.

1. Geological and Geophysical Constrains

1.1. Mantle Convection

[2] When loaded, mantle material undergoes elastic deformation and viscous motion. If the loading persists for a long time (more than a few hundred thousand years), the viscous strain will exceed much the elastic component, and

mantle material can be regarded as a viscous fluid. The heat flux q from the mantle averages 90 mW m^{-2} [*Schubert et al.,* 2001]. The temperature at the core-mantle boundary should be $T_1 \sim 4000$ K because it is higher than the melting point of iron with admixtures of light elements at corresponding pressures ($T \sim 3500$ K) and lower than the melting point of silicates ($T \sim 4500$ K).

[3] If convection were absent in the mantle, deep heat would be removed solely through the conductive heat transfer with a flux equal to $q_0 \approx kT_1/H$ (k is heat conduction and H is the mantle thickness). With $k \approx 6$ W (mK)$^{-1}$, $T_1 \approx 4000$ K and $H = 3000$ km, this yields $q_0 \approx 8$ mW m^{-2}, which is about ten times smaller than the observed value.

Therefore, the mantle is involved in vigorous thermal convection enhancing the heat removal process.

[4] A high gradient of density in the D" layer is due to an abrupt increase in the concentration of heavy components at the core-mantle boundary. This hinders much convection inside this layer that could make the temperature distribution smoother. Accordingly, the temperature change across this layer can amount to 1000 K with mantle heated from below and 2000 K with mantle heated by its inner radioactive sources. If the middle of the D" layer is taken as an effective lower boundary of the origination of plumes and ascending mantle flows, the temperature difference responsible for the heating of the mantle should be $\Delta T \approx 3000$ K. Thermal convection arises only if an ascending material element remains invariably lighter and thereby hotter than the surrounding mantle. However, due to decompression, hot material ascending from the mantle base to the Earth's surface cools by about $\Delta T_{\text{ad}} = 1200$ K. Therefore, mantle convection is due not to the total temperature difference but only to its potential (superadiabatic) part $T_{\text{p}} = \Delta T - \Delta T_{\text{ad}}$. This yields the value $T_{\text{p}} = 1800$ K for the potential temperature of mantle plumes.

[5] The effectiveness of heat transfer in a convective layer is characterized by the Nusselt number Nu, equal to the ratio of the convective heat flux q to the conductive heat flux q_0 caused by the potential temperature difference: $q_0 = kT_{\text{p}}/H \approx 3.6$ W $(\text{mK})^{-1}$. To determine the convective heat flux, we find the conductive heat flux due to the temperature difference in the lower part of the D" layer and the adiabatic difference (the overall difference $\Delta T = 2200$ K): $\Delta q = kT_{\text{p}}/H \approx 4.4$ W $(\text{mK})^{-1}$. Subtracting this value from the observed heat flux, we obtain that the convective heat flux from the mantle amounts to 85 W $(\text{mK})^{-1}$. As a result, the Nusselt number in the Earth's mantle is $\text{Nu} = q/q_0 \approx 85/3.6 \approx 24$.

[6] As is known from the theory of convection in a layer heated at its lower boundary [*Schubert et al.,* 2001], the convective velocity V is connected with the Nusselt number through the relation $V \sim 8(k/H)\text{Nu}^2$, where k is the diffusivity defined as the thermal conductivity divided by density and specific heat; with $k = 10^{-6}$ m^2 s^{-1}, this formula yields a value of about 5 cm yr^{-1} for the velocity of circulation and mixing of mantle material.

[7] Apart from the Nusselt number Nu characterizing the effectiveness of convective heat transfer, another important quantity is the Rayleigh number Ra characterizing the intensity of convection (see formula (8) in Appendix): $\text{Ra} = \alpha g \Delta T H^3/(k\nu)$. Here, α is the thermal expansion coefficient, g is gravity, ν is the kinematic viscosity (defined as dynamic viscosity divided by density), and ΔT is the superadiabatic temperature difference (the potential temperature). Setting in the mantle $\alpha = 2 \cdot 10^{-5}$ K^{-1}, $\nu = 3 \cdot 10^{18}$ m^2 s^{-1} and $\Delta T = 1800$ K, we obtain $\text{Ra} = 2 \cdot 10^6$. In the convection theory, the Rayleigh and Nusselt numbers are interrelated through the approximate formula $\text{Nu} = 0.2\,\text{Ra}^{1/3}$; hence, setting $\text{Ra} = 2 \cdot 10^6$, we obtain $\text{Nu} = 25$ in agreement with the estimate obtained from the observed heat flux of the Earth.

[8] In energy terms, the Earth can be compared with a heat engine in which the heat is permanently radiated into the space and is partially converted into the energy of convective motion. Presently, it is believed that four main sources contribute to the heat flux of the Earth: radioactive decay in the crust ($\sim$15%), solidification heat of the growing inner core ($\sim$10%), radioactive decay in the (predominantly lower) mantle ($\sim$50%), and heat of the cooling Earth ($\sim$25%) that was accumulated during its formation [*Schubert et al.,* 2001]. This balance corresponds to an Earth's cooling rate of 80 K Gyr^{-1}.

[9] Previously, mantle flows were sometimes attributed to chemical, rather than thermal, convection associated with sinking heavy components of mantle material (e.g. remaining iron) and ascending light components (e.g. liberated from the core). Hot light material rises during thermal convection. Cooling at the Earth's surface, this material forms the thickening lithosphere, which sinks when it becomes heavier than the hot mantle material. However, if the lower density of the rising material were due to its chemical composition rather than its higher temperature compared to the surrounding mantle, magma of mid-ocean ridges would have room temperature, which contradicts data of observations. Moreover, a lithosphere formed by this lighter material would not have thickened with time and would not have sunk into the mantle. As mentioned above, thermal convection is due to the nonzero potential temperature of the ascending material. Calculations of the global heat flux distribution also indicate that mantle density inhomogeneities (apart from the continental lithosphere) determined from seismic tomography data are of thermal, rather than chemical, origin because they are consistent with the observed anomalies of the terrestrial heat flux. If a negative density anomaly were of chemical origin, a higher heat flux would not be observed at the corresponding point of the Earth's surface.

1.2. Origin of the Asthenosphere and Lithosphere

[10] The mantle material viscosity depends very strongly (by an exponential law) on temperature and pressure. Pressure increases monotonically with depth (the pressure gradient is proportional to density). In the absence of convection, the temperature rise and viscosity variation with depth should also be monotonic. However, during developed mantle convection, the mantle heat transfer in regions occupied by vertical flows is mainly convective, the temperature gradient being there relatively small. Consequently, temperature in these zones rises with depth solely due to adiabatic heating of material associated with its compression in descending mantle flows (and, correspondingly, cooling in ascending flows).

[11] The vertical heat transfer is purely conductive in the upper and lower boundary layers (where velocities are nearly horizontal). However, because the heat flux in a stationary state should be depth independent, the Earth's temperature first increases very rapidly with depth, from 0°C to $\sim$1400°C in the upper 100 km, after which its gradient drops to a value smaller than 1°C km^{-1}. As a result, the upper cold layer of the mantle (lithosphere) is highly viscous and is underlain by a low viscosity asthenosphere. The boundary of these two

layers is formally associated with the solidus temperature $1300°C$. At greater mantle depths, the effect of slow temperature rise is dominated by the pressure increase, and the mantle viscosity increases with depth by one or two orders. A lower conductive boundary layer with a high temperature gradient arises at the mantle base due to heat flux from the core. As a result, viscosity in the D" layer, as in the asthenosphere, drops and this leads to instability of flows and generation of plumes in this layer. Possibly, the viscosity in the lower 1000 km of the mantle (but above the D" layer) is higher than the value presently accepted.

[12] Due to strong lateral variations in temperature, both lithosphere and asthenosphere vary in thickness. Hot mantle material reaching mid-ocean ridges solidifies, after which it moves horizontally, further cooling and forming the thickening oceanic lithosphere. As the lithosphere becomes heavier, the ocean floor deepens. On the contrary, the underlying asthenosphere becomes thinner away from the ridges and disappears completely near subduction zones. In continental regions, the asthenosphere can only exists in the form of isolated lenses.

[13] A fundamental distinction exists between the oceanic and continental lithosphere. The oceanic lithosphere is involved in the convective circulation of the upper mantle material and, for this reason, its lifetime at the surface is no more than 200 Myr. The continental lithosphere is frozen up to a continent from below, making with it a coherent structure due to mass transfer and becoming lighter due to phase transformations. Therefore, the continental lithosphere (adjacent to the base of the continental crust) exists for more than 3 Gyr. Since the mantle beneath continents (at depths of about 200 km) is $100–200°C$ colder than beneath oceans, the continental lithosphere is much thicker than the oceanic lithosphere. Accordingly, asthenospheric lenses beneath continents arise generally in places where the mantle is heated by hot plumes intruding from below.

1.3. Tectonics of Oceanic Lithospheric Plates

[14] The range of mantle temperatures extends from $0°C$ at the Earth's surface to more than $4000°C$ at the core-mantle boundary, and the pressure increases with depth by six orders of magnitude. Under these conditions (taking into account phase transformations, melting and solidification), the viscosity can vary in the mantle by more than 20 orders, from 10^3 Pa s in basaltic magmas to 10^{26} Pa s in the cold lithosphere.

[15] With such a high viscosity, the oceanic lithosphere (like solidified tar) is fairly brittle, albeit capable of experiencing very slow viscous deformation. When subjected to a variable stress, the lithosphere is broken into a system of horizontally moving plates. When the lithosphere becomes sufficiently heavy, it starts sinking into the mantle in subduction zones. The subducting lithospheric plates (slabs) heat and dissolve in the mantle. Recent tomography data visualize not only the slabs but also their nondissolved remnants in the lower mantle and at its base.

[16] Strictly speaking, the idea of rigid lithospheric plates is incompatible with their bending in subduction zones.

When elastically deformed, the lithospheric plate cannot bend even by a few degrees (an elastic tangential stress cannot exceed the one thousandth of the shear modulus) and should finally break up. Oceanic plates do not break in subduction zones because their material under shear stress conditions becomes ductile. Therefore, the classical tectonics of rigid lithospheric plates is only applicable to horizontal plate movements and small bending associated with underthrusting or overthrusting. At mid-ocean ridges, the ascending mantle flow easily turns to assume a horizontal direction because its material did not solidify as yet. In subduction zones, a change in the movement direction of a plate is associated with a loss of rigidity under conditions of higher stresses.

[17] Plate tectonics is understood, in a narrow sense, as kinematics of quasi-horizontal movements of rigid brittle lithospheric plates including overthrusts, underthrusts, and motions along transform faults, as well as processes of folding and brittle fracture at plate contacts. This theory is deficient in that it does not examine driving forces or treats them as external factors.

[18] The plate tectonics understood in a broad sense is the theory of origination and evolution of lithosphere, including its kinematics and dynamics. Oceanic lithosphere in this theory arises self-consistently as a result of the solution of equations of thermal convection in the mantle with temperature dependent viscosity. This theory describes numerically the processes of the solidification of magma at mid-ocean ridges, seafloor spreading, thickening of the lithosphere with age, and its sinking in subduction zones. However, the theory has not been accomplished as yet. For example, we fail so far to get a self-consistent description of breakage of the lithosphere into plates with the formation of transform faults.

[19] The viscosity of mantle material can be represented as an exponential function of temperature and pressure and as a power function of tangential stress: $\eta = A\sigma_2^{1-n} \exp[(E + pv)/R(T + \theta)]$, where E is the activation energy, v is the activation volume, R is the gas constant, A is a prescribed constant, θ is an additional constant chosen from a given value of the effective viscosity of the cold oceanic lithosphere $\eta|_{T=0} < \infty$, and σ_2 is the second invariant of the viscous stress tensor, $\sigma_2 = (\sigma_{ij}\sigma_{ij})^{1/2}$.

[20] As mentioned above, the full range of the viscosity variation in the mantle encompasses more than 20 orders of magnitude. Without regard for melting processes, this range amounts to 5–8 orders. The exponent n in a dislocation creep model is usually taken equal to 3. The ideas of mantle convection and oceanic lithospheric plates spreading away from mid-ocean ridges and sinking into the mantle in subduction zones were formulated by the 1970s and replaced notions of a solid mantle with separate magma sources. Plate tectonics arose first as a kinematic theory. However, it was found out in the 1980s that constraints on properties of the oceanic lithosphere (except its breakup into plates) can be directly gained from the theory of mantle convection incorporating the temperature dependence of viscosity. Thus, the theories of mantle convection and oceanic lithosphere are presently united into the general theory of mantle convection with variable viscosity.

[21] The above concept of mantle convection with oceanic lithosphere has been developed over the last three decades by geologists, geophysicists and geochemists up to the stage of its numerical description in terms of mathematical modeling [*Anderson,* 1989; *Fowler,* 1996; *Schubert et al.,* 2001].

1.4. Tectonics of Floating Continents

[22] Purely continental plates are absent in maps showing lithospheric plates. So it is assumed that continents are light inclusions frozen up into the oceanic lithosphere which passively drift together with the latter and have no effect on global processes. Although this model of the Earth became the central model in studies and interpretations of regional processes on a time scale of up to 100 Myr, it fails to explain longer-term history of the Earth, treating recent global structures of the Earth as an unpredictable result of chaotic mantle convection.

[23] Likewise, the theory of lithospheric plate tectonics fails to account for such phenomena as the assemblage and breakup of supercontinents, the causes for formation and long-term existence of continental lithosphere and roots of continents, the mechanism of oblique underthrusting of oceanic lithosphere beneath continents, and others.

[24] A new geodynamic model including mechanical and thermal coupling of mantle with floating continents has been developed in [*Trubitsyn,* 2000, 2004; *Trubitsyn and Mooney,* 2002; *Trubitsyn and Rykov,* 2000, 2001; *Trubitsyn et al.,* 2003]. The lifetime of a continent (∼3 Gyr) is much longer than the lifetime of a lithospheric plate. The Earth's surface temperature (∼300 K) being much less than the melting point of mantle material (∼1500 K), continents can be compared with huge rafts floating in a varying ice field. "Ice" temporarily frozen up on the sides of continents corresponds to the oceanic lithosphere at passive continental margins. The "ice" frozen up from below for a long time corresponds to the continental lithosphere. However, as distinct from the ordinary ice (which is lighter than water), the heavy oceanic lithosphere at active continental margins can sink into the mantle, and its material is involved in the convective mantle circulation. Due to thermal screening, the heat flux from the mantle in continental regions is three times as small as in oceanic regions. Therefore, floating continents change basically the whole evolution of thermal mantle convection. In order to describe the coupling between viscous mantle and rigid continents, the system of the mantle convection equations is complemented with the equations of motion of rigid continents (in a spherical model, these are the Euler equations of dynamics of a solid). Continents of varying thickness and arbitrary sizes and shape can move on the surface, rotate under the action of viscous forces applied to their submerged surfaces, and collide. Unknown are the velocities of mantle flows and continents, as well as the temperature in the mantle and the continents. Solution of these equations also provides the distributions of viscosity, viscous stress tensor, and heat flux, the seafloor topography, the gravitational field, and movements of continents.

[25] The Earth was sometimes compared with a heat engine, with the mantle being similar to a boiler, and oceanic plates, to connecting rods. However, without actuator valves, the heat engine cannot ensure an ordered motion. It was found out that heat-screening continents play the role of floating valves that redistribute the heat flux in the mantle and control the geologic history of the Earth. Within the framework of the mantle convection model with floating continents, global processes in the Earth are not a result of chaotic convection (as was recently believed in tectonics of lithospheric plates) but can be consistently calculated and thereby predicted. The evolution of the mantle-continents system develops between two quasi-equilibrium states: the state in which each continent is close to adjacent downwelling mantle flow and the state in which all continents are assembled into one supercontinent above a system of mantle downwellings. One might say that continents actually control the evolution of mantle convection and the tectonics of lithospheric plates.

[26] The present study, complementing the papers [*Trubitsyn,* 2000; *Trubitsyn et al.,* 2003] and presenting modeling results for mantle convection with two floating continents, treats in greater detail the formation of a mantle plume system beneath a supercontinent and gives analysis of uplift and subsidence of continents with related fluctuations in the ocean level incorporating data of recent reconstructions of the continental drift.

2. Formation and Breakup of Supercontinents

[27] Within the framework of the theory of mantle convection and tectonics of lithospheric plates, both assemblage of continents and breakup of supercontinents were accounted for by the nonstationarity of convection. At large Rayleigh numbers, thermal convection is nonstationary, the shape and sizes of cells permanently vary, and mantle flows are reorganized. For this reason, it was believed that two or more continents can randomly converge toward a mantle downwelling, forming a supercontinent that can in turn break up if a large mantle upwelling arise beneath it.

[28] However, numerical modeling showed that the processes of continental drift and formation and breakup of supercontinents are not chaotic but obey laws of the tectonics of floating continents. Continents are coupled with the mantle through nonlinear thermal and mechanical interactions, so that not only the mantle affects continents but also the continents can reorganize mantle flows.

[29] To gain constraints on the coupling between mantle and continents, we examined a simple 2-D Cartesian model of two continents floating on a highly viscous mantle. We took the simplest model of the type $\eta = 10^{23} \exp(-4.6T + 0.9H)$ Pa s, in which the viscosity decreases by two orders with increasing temperature and increases by 2.5 times with increasing pressure. The effective Rayleigh number was specially taken very small (Ra = $2 \cdot 10^4$) to make the convection intensity low, with the Nusselt number Nu = 6. Convection in this model is stationary and laminar, so that no plumes can arise in this model in the absence of continents.
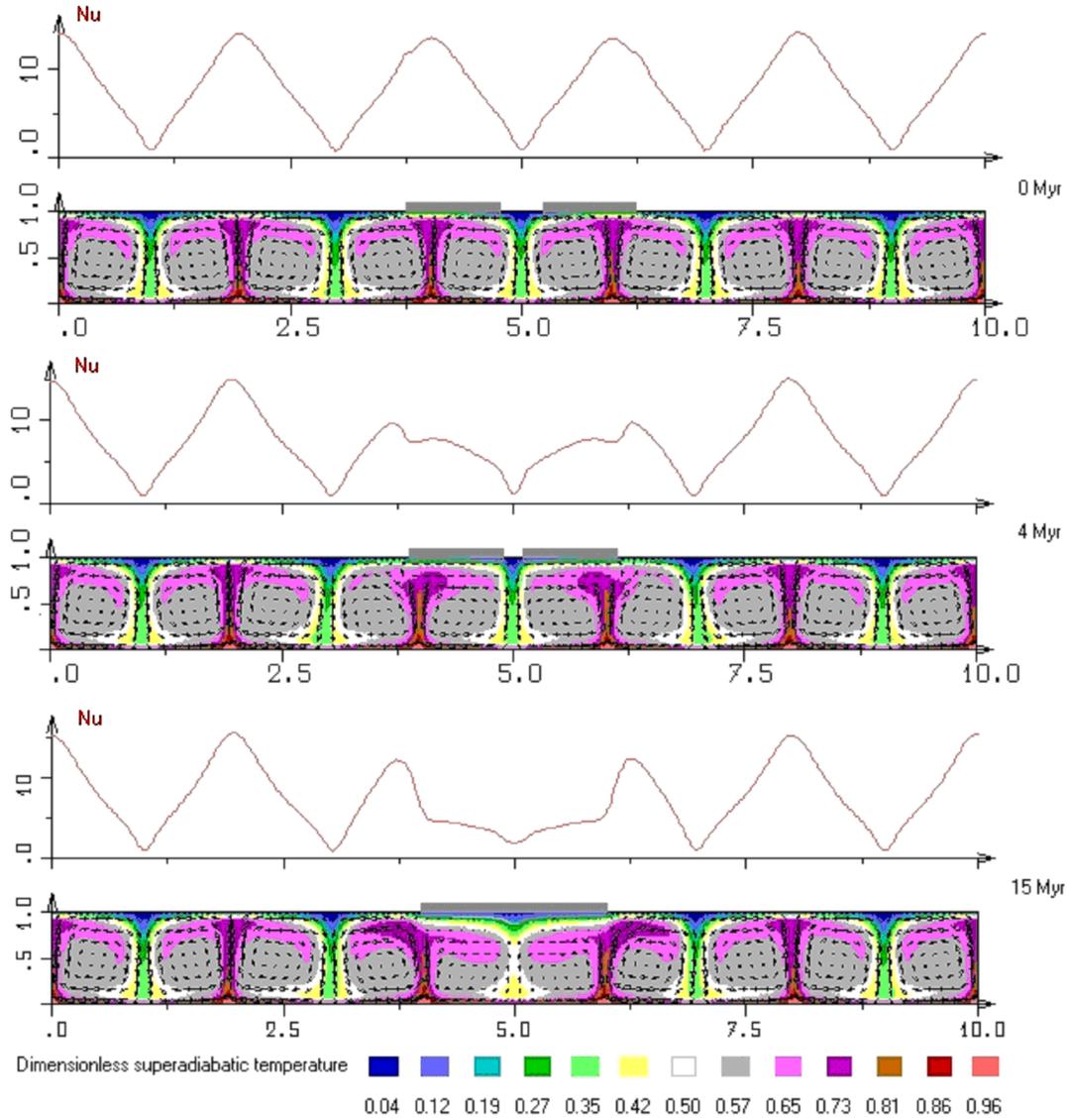
Figure 1. Structure of mantle convection calculated at the stage of the aggregation of continents into a supercontinent. The dimensionless superadiabatic temperature is shown in color. The arrows are mantle flow velocities. The longest arrow corresponds to a velocity of 4 cm yr^{-1}.

[30] After the thermal convection in the mantle had attained a steady-state pattern, two freely floating continents were placed on the mantle. They were modeled by two solid floating plates 300 km thick. These plates can be interpreted as incipient primary continents consisting of continental crust coupled with the oldest strong thin continental lithosphere. Numerical solution of equations of mass, heat and momentum transfer in viscous mantle and Euler equations for the solid continents provides a description for the evolution of the mantle–continents system including the origination, growth, and further evolution of the continental lithosphere. The equations of mantle convection and boundary conditions, as well as a method of their solution, are described in Appendix.

[31] Figures 1–5 present results of the calculation of evolu-

tion of the mantle–continents system. As mentioned above, we have Nu ≈ 25 for real mantle convection in the present Earth. Therefore, these results are of qualitative nature. When extrapolating results of this model to convection in the present mantle, one should take into account that ascending and descending flows will be narrower and more nonstationary for a larger Rayleigh number. In such extrapolation, numerical values of temperature and flow velocity variations can be converted using the known formulas of the theory of parametric convection.

[32] Mantle flow velocities are shown in the figures by black arrows whose maximum length corresponds to a value of about 4 cm yr^{-1} (upon conversion to Ra = 10^6). The dimensionless temperature distribution is characterized by various colors. Hot mantle upwellings are shown by red,
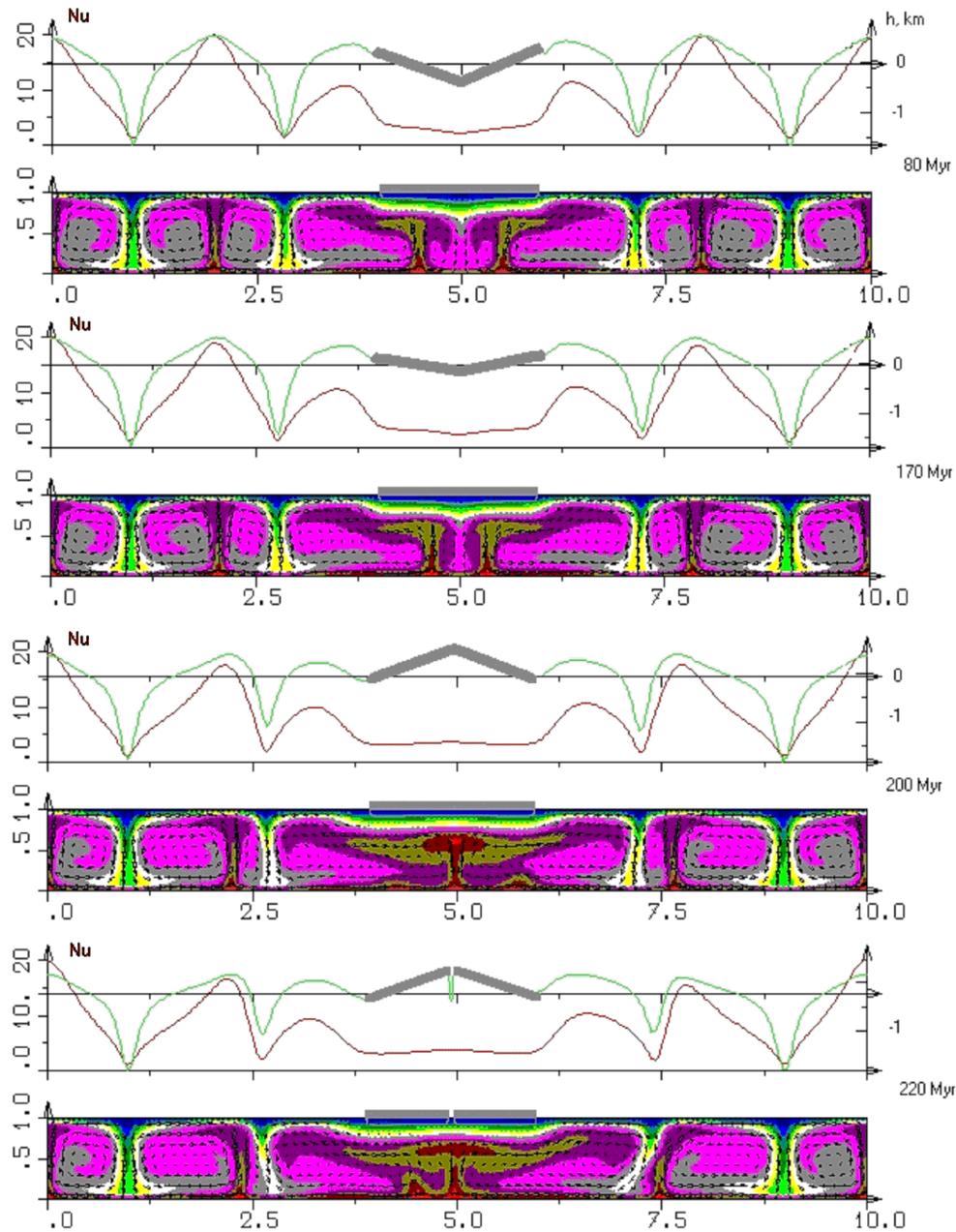
Figure 2. Evolution of mantle convection at the supercontinent stage.

whereas the cold highly viscous oceanic lithosphere sinking in subduction zones and transformed into cold mantle downwellings is shown by blue. The continents are represented by gray segments. Reddish brown colors the calculated distribution curve of the dimensionless mantle heat flux Nu(x). The green line is the calculated seafloor topography.

[33] Figure 1 presents the structure of mantle convection calculated at the stage of assemblage of continents into a supercontinent. At the time moment $t = 0$, when the continents were placed into the mantle, convective cells had a regular symmetric shape. The heat flux is elevated (about 200 mW m^{-2} after the conversion to Ra $= 5 \cdot 10^6$) above

mantle upwellings and reduced above downwellings. The continents seat on both sides of the middle mantle downwelling and are not underlain by a continental lithosphere at the initial time moment.

[34] Calculation of the evolution of the mantle-continents system showed that the continents start moving toward the downwelling under the action of tangential viscous forces. By the time moment $t = 15$ Myr, each continent moved for about 300 km at an average velocity of about 2 cm yr^{-1}. By this time, the continents converged and formed a supercontinent. Since the heat transfer mechanism in the continents is purely conductive, they hinder the removal of heat from the
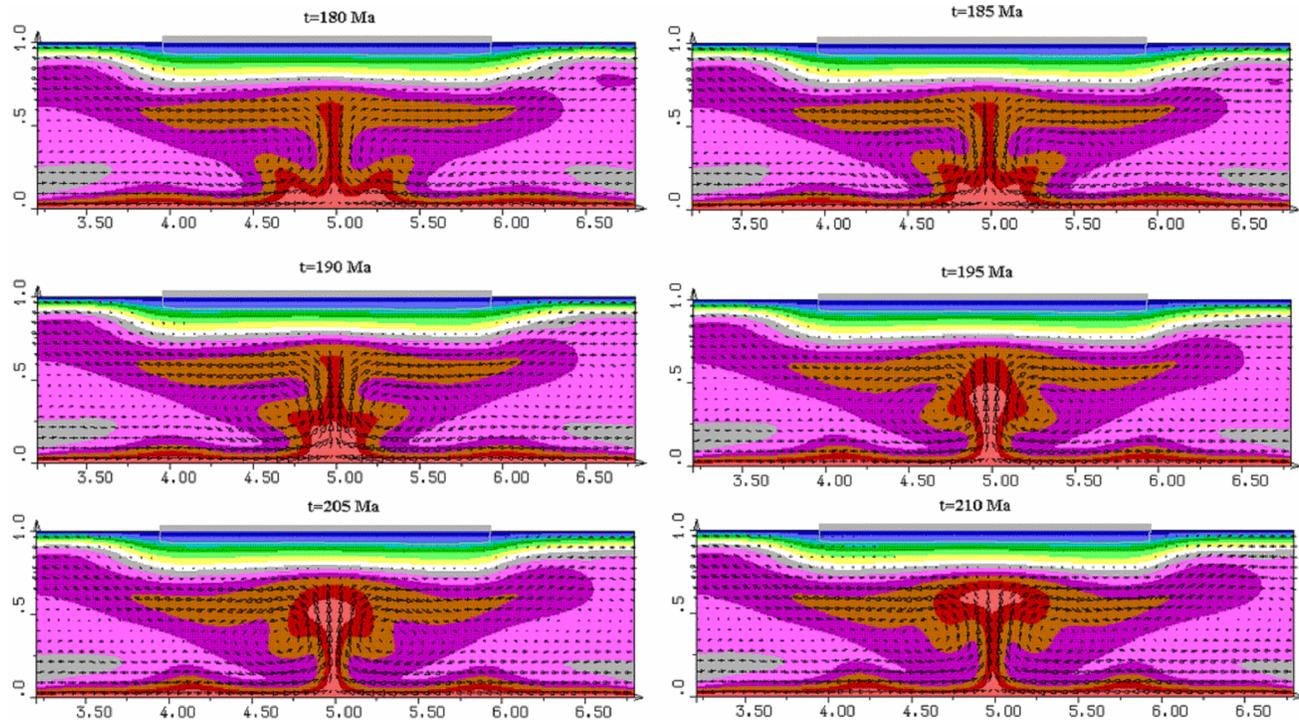
Figure 3. Origination and evolution of a mantle plume beneath the supercontinent.

mantle. As seen from Figure 1, the heat flux through the supercontinent (about 30 mW m^{-2}) is nearly six times smaller than in ridge zones. The mantle material under the supercontinent is still cold and heavy and continues to descend, compressing the supercontinent.

[35] Further calculation of the mantle convection evolution at the supercontinent stage is illustrated in Figure 2. Due to the heat screening effect, the cold mantle material beneath the supercontinent starts heating and thereby becomes lighter. Approximately after 200 Myr (with other values of parameters, after 400 Myr [*Bobrov et al.,* 1999]), the downwelling beneath the supercontinent gives way to a giant ascending mantle flow withdrawing heat from the mantle base. However, as seen from Figure 2, this hot mantle upwelling increases the heat flux above the supercontinent still insignificantly due to the heat-screening effect of the crust and continental lithosphere.

[36] Figure 3 shows the region under the supercontinent on a larger scale. Isolated floating-up diapirs periodically arise at the center of the wide mantle upwelling; when detached, they are transformed into separate hot spots (thermals). Note once more that, in the absence of continents, plumes cannot arise in this model with a small Rayleigh number (see Animation 1 in online version of the paper for more details of plume evolution).

[37] Figure 4 shows the convection structure and the position of the continents at the stage of their divergence after the breakup of the supercontinent. The continents diverge in opposite directions from the central hot mantle flow. Although the mantle beneath the continents is very hot, the heat flux crossing the continents is still small (40–

50 mW m^{-2}) because the heat from the hot mantle plume did not reach the Earth's surface at this time.

[38] As the continents continue diverging, the hot mantle upwelling is enhanced. Accordingly, the heat flux from the mantle also increases, attaining maximum values of more than 200 mW m^{-2}, but does not exceed 20–30 mW m^{-2} above the continents because they hamper the heat transfer.

[39] At the time moment $t = 290$ Myr, a structure of the Atlantic Ocean type arises between the continents. A mid-ocean ridge with a high heat flux is observed in its axial zone. The colder region of the mantle (corresponding to the oceanic lithosphere) thickens away from the ridge. On the other side of the continents, there arose a structure of the Pacific Ocean type along with subduction zones, where the oceanic lithosphere plunges into the mantle beneath the continents.

[40] The structure that arose by the time moment $t = 340$ Myr is hypothetically conceivable as a future Atlantic Ocean when its greatly extended oceanic lithospheric plates will start sinking at the margins of Europe and North America. From this time, the Atlantic Ocean can be similar to the present Pacific Ocean (see Animation 2 in online version of the paper for more details).

[41] The calculated models are a result of numerical experiments based on the solution of classical differential equations of energy, mass, and momentum transfer. These experiments show that, during convection, two plates floating over a heated viscous fluid must periodically converge above a descending flow, after which they must inevitably diverge. The time between their assemblages depends on the viscosity and temperature of the fluid. Taking into account results of
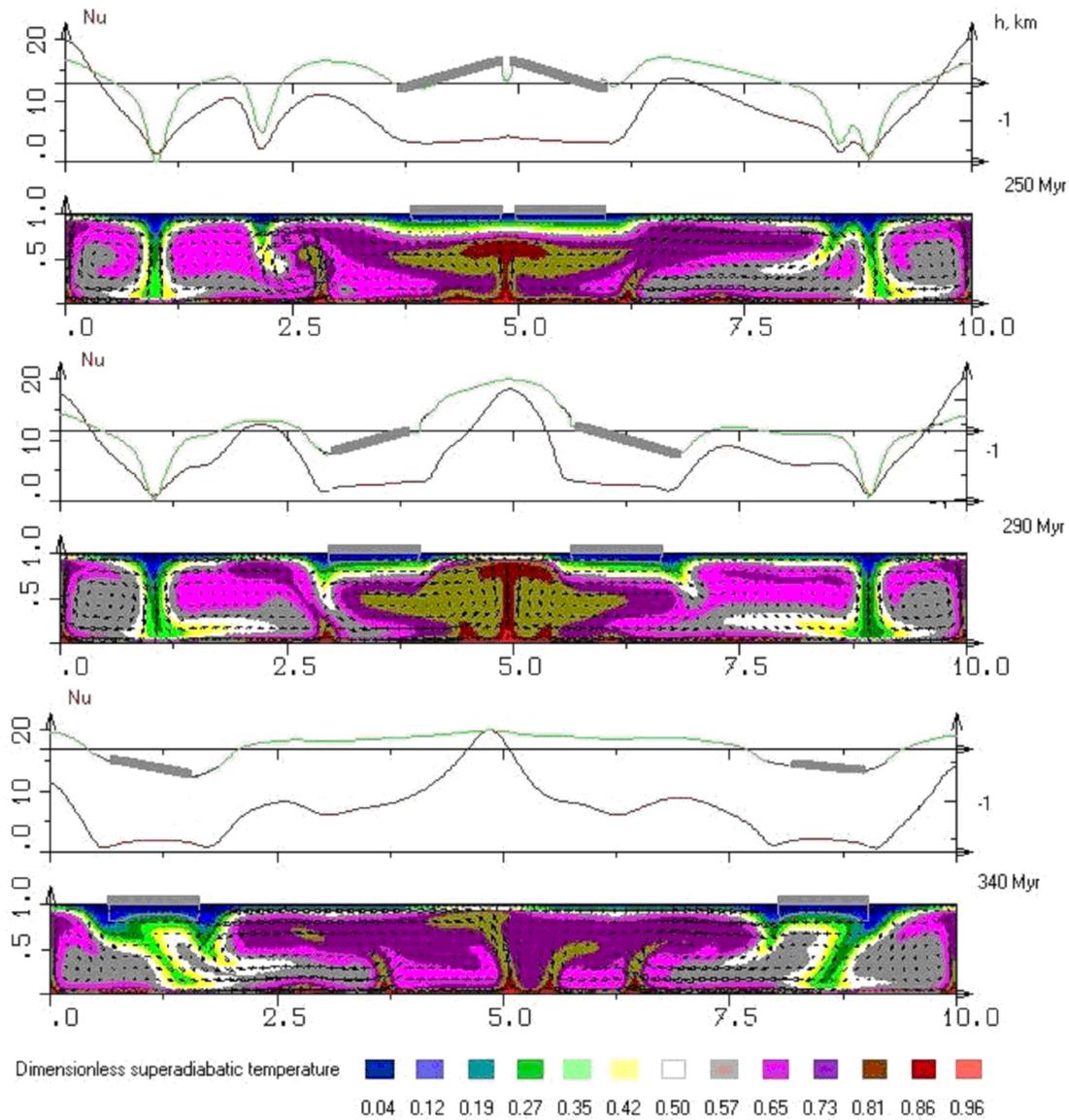
Figure 4. Structure of mantle convection calculated at the stage of dispersal of continents.

similar 3-D model calculations presented in [*Trubitsyn and Rykov,* 1995] and converting them to mantle parameters, it may be supposed that the time intervals between formations of supercontinents amounted to 0.5–1.0 Gyr.

3. Formation and Evolution of the Continental Lithosphere

[42] The formation and evolution of the continental lithosphere can be easily conceived by analyzing the above modeling results obtained for the evolution of the mantle floating continents system. As seen from Figure 1, two continents in the form of foreign plates were placed onto the mantle at the initial time moment. Later, in the process of evolution (see Figure 2 and Figure 4), significant changes took place in the mantle due to the effect of the continents and, in particular, cold (and thereby highly viscous) stagnant regions arose in the vicinities of the continents. Each of these regions moves together with its continent.

[43] The distribution of the dimensionless temperature is shown in color in Figures 1–4. With an overall temperature difference of about 4000 K, the temperature shown by green is about 1200 K. Assuming that this value corresponds to the lower boundary of the lithosphere, the thickness of the continental lithosphere in such a simplified model should amount to a few hundred kilometers. As noted above, a higher intensity of thermal convection should result in thinner thermal boundary layers and, therefore, in a thinner continental lithosphere. Conversion based on the theory of parametric
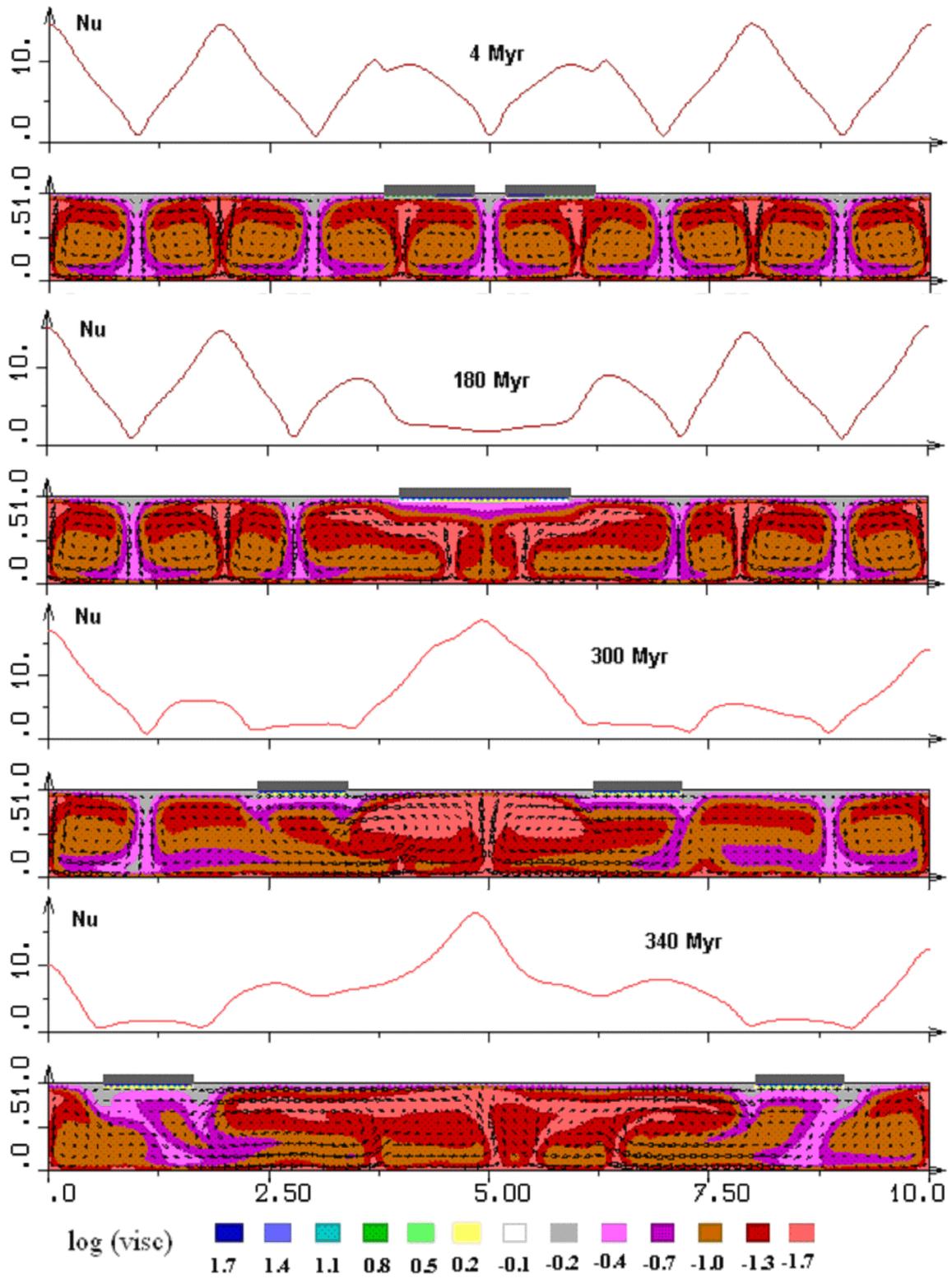
Figure 5. Viscosity distribution evolution in the mantle with floating continents. The logarithm of relative viscosity is shown in color. The maximum viscosity variation reaches 3.5 orders.

convection shows that with Ra $\approx 10^6$ the thickness of the oceanic lithosphere is about 50 km and that of the continental lithosphere is on the order of 200 km (400 km with roots).

[44] Figure 5 shows mantle flow velocities against the mantle viscosity distribution (blue and red correspond to high and low viscosity values).

[45] The numerical experiments presented in this paper show that the continental lithosphere varies with time in thickness and configuration. The continental lithosphere thickens when a continent moves into a colder mantle region and thins when it is in a hotter region. Since the mantle temperature significantly varies over the length of the continent, the shape of the continental lithosphere should also change. In particular, the lithosphere thins where it overlies an approaching hot plume and deep continental roots develop above anomalously cold mantle regions.

[46] The continental lithosphere is seen most distinctly in Figure 4, showing the situation after 300 Myr. By this time, the continents diverged for a distance of about 10,000 km and an ocean with a system of hot mantle upwellings formed between them, where the mantle heat flux is anomalously high. A cold, highly viscous continental lithosphere grew under each continent. It has an irregular shape, is somewhat thinned above ascending plumes, and has deep roots above descending cold flows. Thus, in terms of a fairly simple model, a continental lithosphere arises in a self-consistent way, without any additional assumptions. A physical interpretation of this phenomenon consists in that continents are permanently drawn by descending mantle flows into the coldest mantle regions. As a result, highly viscous zones arise under the continents and promote the formation of the continental lithosphere (see Animation 3 in online version of the paper for more details of the evolution of mantle convection with floating continents and varying viscosity).

[47] Oceanic lithospheric plates resemble ice floes temporarily freezing to a floating body on its sides. The continental lithosphere forms when mantle material is frozen to the floating continent from below. The lower boundary of the lithosphere is defined by the partial melting point of mantle material (about $1200°C$). Therefore, the continental roots can be compared to icy outgrowths whose interior is a residue from smelting of the crust.

[48] However, the oceanic lithosphere basically differs from ice. As distinct from floating floes, cooling oceanic lithospheric plates become heavier than the underlying mantle, sink into the mantle, and gradually mix with its material. Likewise, the primary continental lithosphere could have mixed in the mantle. However, due to a low temperature and a high viscosity, mantle flow velocities under a continent are lower than horizontal velocities of oceanic plates. Therefore the continental lithosphere has a chance to be reworked by repeatedly arising plumes. Moreover, it is permanently subjected to metamorphic transformations promoted by fluid flows. As a result, the continental lithosphere becomes lighter and rigid and is "welded" to the continent. Note also that an older lithosphere is stronger. Possibly, it is a strong lithospheric substrate of large continents that prevents them from being crushed by oceanic plates.

[49] For convenience of analysis, the continental lithosphere can be conventionally subdivided into two parts (although no boundary separates them): the upper older (metamorphized and lighter) part and the lower younger part that differs from the underlying mantle by a higher viscosity (due to a lower temperature). Whereas the older subcrustal continental lithosphere is more stable, the lower younger continental lithosphere permanently varies. In higher temperature regions, it softens and can be carried away by mantle flows. This can account for great variations in the thickness of continental lithosphere observed both between different continents and within a continent.

[50] We should note that the results presented in our paper do not take into account processes of transformations and transportation of chemical elements in the continental crust that lead to its strengthening. A floating continent includes not only the crust but also the oldest strong stable part of the continental lithosphere. Numerical experiments show the origination and evolution of the lower younger continental lithosphere beneath continents. The continental lithosphere first automatically arises as an anomalously cold mantle layer frozen to the base of a continent and moves together with the continent due to an anomalously high viscosity. However, with a temperature drop by about 200 K, the viscosity of the continental lithosphere increases by only two orders. This is insufficient in order to completely inhibit the mixing of the continental lithosphere with mantle flows. In the real Earth, the formation of the continental lithosphere is accompanied by material differentiation.

[51] The mixing of the lithosphere with the mantle is also hindered by buoyancy. The higher viscosity of the cold lithosphere gives it a chance to "dry" and become strong to such a degree that, along with continental roots, it remains attached to floating continents for a few billion years.

4. Global Sea Level Variations

[52] Data on shallow-water deposits and many other features indicate that variations in the sea level have existed throughout the geologic history of the Earth. A wide spectrum of sea level variation cycles of various periods has been established [*Turcotte and Schubert*, 1982; *Veil et al.*, 1977]. Global sea level variations can be due to climatic processes, in particular, formation and thawing of glaciers, when water is first gathered on continents and then comes back into oceans. The ocean water volume can also vary if the amounts of water transported by magma from the mantle and escaping into the mantle in subduction zones are different. These processes change the sea level synchronously for all continents by up to a few hundred meters. Moreover, regional uplifts and subsidences of land due to tectonic processes are possible on each continent and their amplitudes also reach hundreds of meters. These changes can evidently be asynchronous in both continents and their regions.

[53] At present, most unclear is the origin of sea level variations with the longest periods (on the order of a few hundred million years). The curve of such variations obtained by *Veil et al.* [1977] and shown in the middle panel of Figure 6 was constructed from all borehole data available
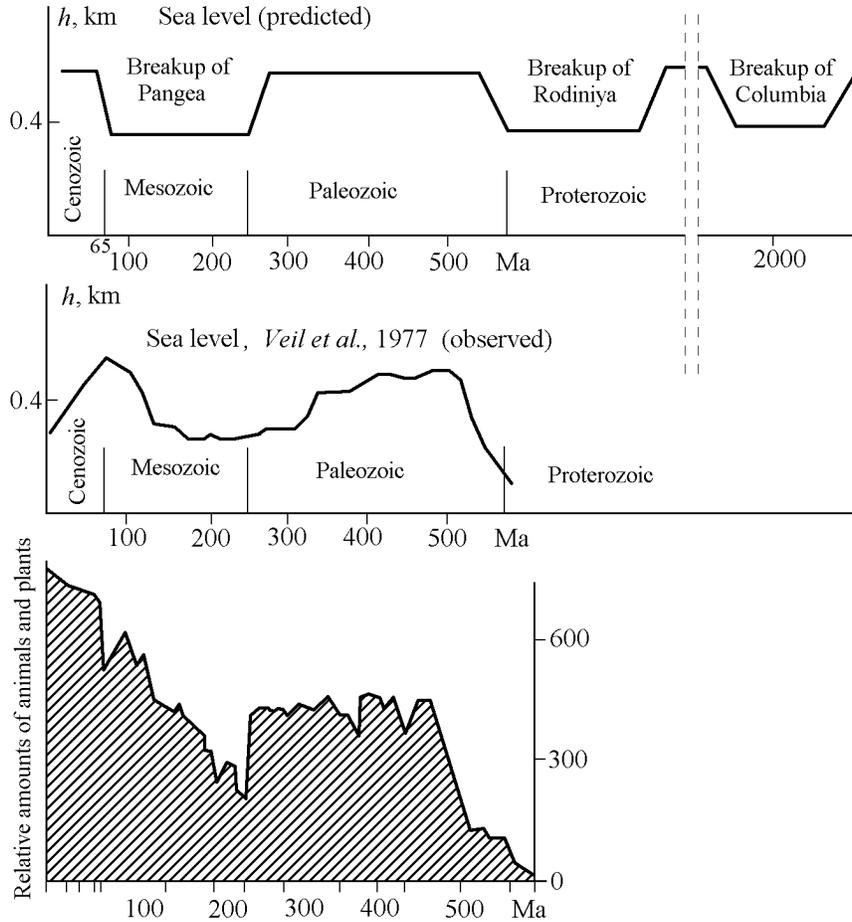
Figure 6. Global variations in the sea level: schematic curve calculated in this paper (upper panel), data of observations [*Veil et al.*, 1977] (middle panel), and curve of the relative amount of species of plants and animals (lower panel; after [*Sepkoski*, 1987]).

at that time; the data were averaged over all continents but characterize mainly North America. Because of long periods of these variations, the sea level was assumed to have varied synchronously for all continents.

[54] We compare the origin of global sea level variations with theoretical inferences of the theory of floating continents, in particular, with the above results derived from a numerical evolution model involving two continents. The green-colored curves in the upper parts of Figure 2 and Figure 4 depict the model topography of the mantle surface with floating continents, (i.e. variations in the seafloor topography $h(x)$ at various time moments).

[55] The sea level coincides with the geoid surface. The topographic inequalities of the geoid are on the order of a few tens of meters, whereas the amplitude of variations in the seafloor topography reaches a few kilometers. Therefore, given a constant ocean water volume, the averaged seafloor surface displaced by an average ocean depth can be taken, in a first approximation, as a sea level.

[56] In the absence of water, the mantle surface topography h_m can be found from the relation connecting the surface height with the vertical stress $h_m = (p + S_{zz})/(\rho_m g)$,

where ρ_m is the mantle density. Here, the pressure p and the stress S_{zz} are calculated in the process of the solution of the convection equations (see Appendix). The topography h of the seafloor covered by water is calculated with regard for the water buoyancy force: $h = (p + S_{zz})/(\rho_m - \rho_w)g$, and its amplitude increases by about one-third. The floating continents rise or subside following the variations in the topography of the mantle surface relative to its average level. The continents jut out above the ocean surface. Variations in their position relative to the average ocean level are controlled by the mantle topography in the absence of water, rather than by the seafloor topography, because a rise in a continent decreases its buoyancy thereby compensating the difference between the topographies of the seafloor and the mantle surface.

[57] The black segments in Figure 2 and Figure 4 show schematically the position of continents relative to the average ocean level (the thickness of the continents is neglected here). Calculations of mantle convection still involve a number of uncertainties, and the goal of this study was to gain qualitative constraints rather than to attain numerical coincidence with results of observations.
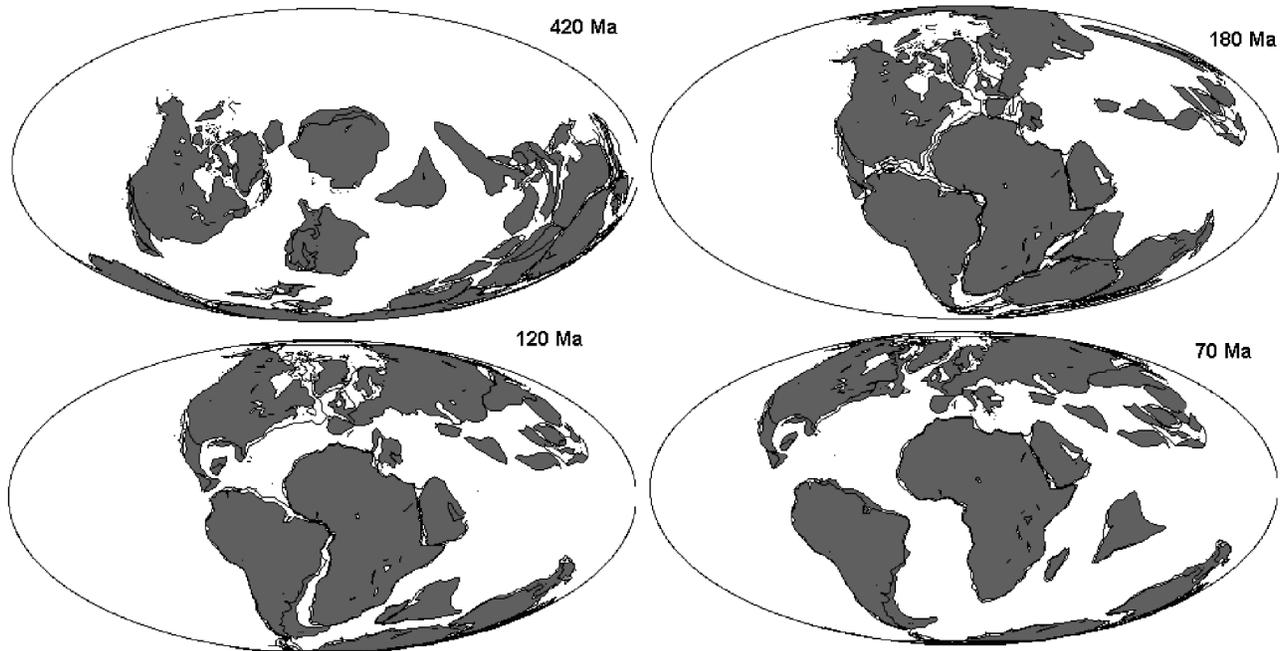
Figure 7. Aggregation and dispersal of continents reconstructed for times of 420, 180, 120, and 70 Ma [*Scotese et al.,* 1988] as an illustration of the Pangea evolution.

[58] In our model, even at the moment when continents are placed onto the mantle, the surface of the latter (and, accordingly, the seafloor) is irregular. The seafloor is up-lifted above mantle upwellings and subsides above down-wellings. Model recalculations with parameters of the real mantle yield a variations in the amplitude of the seafloor topography of about 2–5 km. As is known, the ocean depth averages about 4 km (about 2 km in ridge zones and about 6 km in subduction zones). The model results reveal an in-teresting, previously unknown feature: since the continent sizes are comparable with the wavelength of the convect-ing mantle topography, a solid floating continent does not drift strictly horizontally but experiences variations in its tilts (as a ship at sea). Thus, if Eurasia were in the place of the Pacific plate, the difference between heights of its edges would amount to a few kilometers. Smoothing the topogra-phy of the mantle surface by floating continents decreases, to an extent, their tilts.

[59] As seen from Figure 2, the level of the newly formed supercontinent averages zero because the mantle downwelling attenuated and a new hot flow has not arisen as yet. However, before its breakup, the supercontinent as a whole is uplifted by about 0.5 km, being subjected to a bending force that appears to facilitate its subsequent breakup.

[60] As is evident from Figure 4, which shows the situa-tion arising immediately after the breakup of the supercon-tinent, the mantle flow still uplifts the diverging continents. Figuratively speaking, the continents seem to go down from a hill. Finally, after a large ocean forms between the di-verging continents, they subside to a zeroth level (Figure 4, $t = 340$ Myr).

[61] Thus, our model shows that ocean level fluctuations can be due to not only global variations in the water amount and other possible factors but also due to uplifts and sub-sidences of individual continents and should be regarded as relative, rather than absolute, variations. In a first approxi-mation, various factors responsible for sea level fluctuations can be estimated additively. Note that the position of conti-nents is measured from the mean level of the mantle surface calculated at a given time moment. This allows one to au-tomatically take into account the fact that, when continents rise onto ridges, the displaced water moves toward deep sea trenches, and vice versa, when continents occupy the places of deep sea trenches, water is displaced toward mid-ocean ridges.

[62] This model also shows that variations in the sea level occur synchronously only during the lifetime of a supercon-tinent. In other epochs, different continents diverge at dif-ferent velocities and overlie mantle surface depressions at different times.

[63] The model of mantle convection with floating conti-nents presented in this paper is fairly general and actually independent of initial conditions. Given a different initial configuration, the continents also become, with time, close to a descending mantle flow, as was accepted in our calcula-tions to save the computation time.

[64] In order to compare the modeling results with the real Earth, they should be tied to a concrete time. Paleomagnetic and geologic reconstructions of the history of supergroups-nents can be used for this purpose. Figure 7 illustrates re-constructions of the continental drift before and after the breakup of Pangea [*Scotese et al.,* 1988] at time moments of

420, 180, 120 and 70 Ma. All continents assembled to form Pangea about 250 Ma, and the accomplished cycle amounts to 500–800 Myr.

[65] A giant plume that should have formed beneath Pangea at about 220–250 Ma uplifted the supercontinent and broke it up. For this reason, the sea level relative to Pangea dropped in the Mesozoic. The continents moved to their places near subduction zones at about 100 Ma and the relative sea level rose.

[66] Continents were also uplifted and the relative sea level was low in the Precambrian, during the breakup of Rodinia. Later, in the Paleozoic, when the continents were already located above mantle downwellings and therefore occupied a lower position, the sea level was high. A schematic curve of the sea level variation obtained from our modeling data correlated with time according to [*Scotese et al.,* 1988] is shown in the upper panel of Figure 6.

[67] Overall, the calculated results qualitatively agree with observations, notwithstanding the simplicity of the model. However, a certain shift in phase is observed. According to the Veil curve, the sea level dropped in the Paleozoic and rose in the Mesozoic about 50 Myr earlier. The stratigraphic data used in [*Veil et al.,* 1977] mainly characterize North America. However, the results of the continental drift reconstruction indicate that North America broke off from Pangea at about 175 Ma, whereas South America formed as late as 125 Ma. Consequently, the drift of North America in the Phanerozoic was ahead of other continents. As regards the present time, the present Eurasia is only approaching a subduction zone, the edge of South America overrode a subduction zone, and North America has already moved over a former mid-ocean ridge and this can lead to an uplift of this continent. Therefore, the sea level variations relative to North America (particularly, its western margin) can be some 50 Myr ahead of sea level variations relative to other continents.

[68] It is still unclear whether the sea level drop over the last 50 Myr fixed by the Veil curve is related to an uplift of the western part of the North American continent. The modeling results are only preliminary. On the other hand, the data of observations are averaged over all continents, whereas the model under consideration indicates that continents can be uplifted simultaneously but subsidence periods are specific to each continent.

[69] The interpretation proposed here for the global variation in the sea level on a time scale of a few hundred million years is at variance with the generally accepted ideas. *Turcotte and Schubert* [1982] and *Schubert et al.* [2001] explain a sea level rise in terms of water displaced by the system of mid-ocean ridges. The latter arise due to enhancement of tectonic activity caused by chaotic mantle convection.

[70] The geologic time scale of the Paleozoic, Mesozoic, and Cenozoic has long been used by geologists. One might try to correlate these periods with global variations in the sea level and epochs of great extinctions and revivals of the Earth's organic world. During and after the Rodinia breakup in the Precambrian, continents were uplifted and the primitive life developed very slowly. At about 650 Ma, the ocean rose by nearly 500 m, while the present land level amounts

to about 800 m. Therefore, the ocean flooded nearly half of the land [*Schubert et al.,* 2001; *Turcotte and Schubert,* 1982], and water ruined land protozoans and stimulated the development of new life forms, in particular, the marine animal kingdom prevailing in the Paleozoic.

[71] The ocean receded before the breakup of Pangea, approximately 220–250 Ma. The land area became 1.5 times larger [*Turcotte and Schubert,* 1982]. New lowlands should have been covered by luxuriant vegetation, which stimulated the development of numerous species of giant phytophagous land animals. Finally, at 60–100 Ma (the beginning of the Cenozoic), continents started again subsiding and the sea level rose by a few hundred meters. This resulted in extinction of the land species existing at that time. It is interesting that the sea level variations affected about 50% of land, and great biomass changes were also on the order of 50%. A curve of relative amounts of plant and animal species [*Khain,* 1994; *Sepkoski,* 1987] presented in the lower panel in Figure 7 correlate in part with the sea level variation curve.

[72] As is known, a giant asteroid fell onto the Earth at 65 Ma; it produced a 180-km crater and continental Ir anomalies. This should have led to a climate change and rapid extinction of many species, particularly, the largest ones. Actually, a sharp narrow minimum observed in the curve in Figure 7 exactly coincides in time with the asteroid fall, but the amplitude of the minimum is as low as about 10%. Possibly, the extinction of Mesozoic animals was gradual due to a slow rise in the sea level, and the asteroid brought this process to its accomplishment.

5. Conclusion

[73] Two-dimensional numerical modeling of thermal convection in a mantle with two floating continents has shown that movements of the continents obey certain laws. Continents are coupled with the mantle through thermal and mechanical interactions, so that not only does the mantle affect continents, but the continents themselves reorganize mantle flows.

[74] An inevitable stage of the drift of continents, which are pulled into areas of cold descending convective flows, is their convergence and the formation of a supercontinent. The supercontinent gives rise to the heat-screening effect, causing the heating and decompaction of the underlying cold matter of the mantle. As a result, the descending flow beneath the supercontinent gives way to a giant ascending mantle flow withdrawing heat from the mantle base. Floating-up diapirs and thermals arise periodically at the center of the broad ascending flow beneath the supercontinent. The effect of the ascending flow eventually leads to the breakup of the supercontinent. As parts of the supercontinent diverge, the hot mantle flow is intensified and its heat outflux increases to about 200 mW m^{-2}. However, the heat flux above the diverging continents remains on the order of 20–30 mW m^{-2}. After a certain time, a structure of the Atlantic Ocean type arises between the continents; this structure includes a mid-ocean ridge characterized by high values of the heat flux. The cold mantle region (correspond-

ing to the oceanic lithosphere) thickens away from the ridge axis. Outside this region, a structure of the Pacific Ocean type arises, including subduction zones where the oceanic lithosphere plunges into the mantle under continents.

[75] Numerical experiments have shown that the continental lithosphere changes with time in thickness and shape. The continental lithosphere thickens when a continent is located above a colder mantle region and thins when it is above a hot mantle region. Since the temperature in the mantle varies significantly along a continent, the shape of the continental lithosphere should also vary. Its thickness decreases in regions where a hot plume approaches the lithosphere, and deep roots grow at continents in regions of an anomalously cold mantle. As regards the oceanic parts of lithospheric plates, they can be compared to ice floes temporarily freezing to continental edges.

[76] The continental lithosphere can be tentatively subdivided into two parts: the older and lighter metamorphized upper part and the younger lower part differing from the underlying mantle only by a higher viscosity. The older upper part of the continental lithosphere is more stable, whereas the lower part permanently varies. In softened higher-temperature regions material of the continental lithosphere is washed away by mantle flows. This accounts for great differences in continental lithosphere both between different continents and within each continent.

Appendix A1: Equations of Mantle Convection

[77] Thermal convection in a viscous mantle is described by the distribution of the convective velocity vectors $V_i(x, y, z)$, temperature $T(x, y, z)$ and pressure $p(x, y, z)$. These unknown functions are found by solving the system of momentum-, heat-, and mass-transfer equations

$$\rho dV_i/dt = -\partial p/\partial x_i + \partial S_{ij}/\partial x_j + \rho g \delta_{i3} \ (i = 1, 2, 3) \ , \quad (1)$$

$$dT/dt = \partial(k\partial T/\partial x_i)\partial x_i + H \ , \quad (2)$$

$$\partial \rho/\partial t + \partial(V_i \rho)/\partial x_i = 0 \ , \quad (3)$$

where ρ is the mantle density, g_i is gravity, T is the temperature measured from the adiabatic distribution, k is diffusivity, H is the thermometric density of heat sources, δ_{ij} is the Kronecker delta, and S_{ij} is the deviatoric tensor of viscous stresses:

$$S_{ij} = \eta(\partial V_i/\partial x_j + \partial V_j/\partial x_i) \quad (4)$$

(here η is the mantle viscosity).

[78] Equation (1) means that the acceleration of a fluid element is proportional to the pressure gradient, the resultant of the viscous friction forces, and the gravity force. Equation (2) means that the variation in the temperature of a fluid element is controlled by conductive heat flux and heat generation by internal sources of this element. According to Equation (3), the density variation of the fluid in a given unit volume is controlled by the mass influx from the surrounding fluid.

[79] The inertial terms on the left-hand side of the momentum transfer Equation (1) are on the order of $k\rho/\eta \approx 10^{-23}$ relative to the other terms of the same equation and can therefore be neglected. Using the Boussinesq approximation, we set $\rho = \rho_0(1 - \alpha T)$ in the last, buoyancy-related term of Equation (1) and $\rho = \rho_0$ in all remaining terms of Equations (1–3). The pressure will be measured from its hydrostatic distribution $p(z)$ determined by the condition $\nabla p_0 = -\rho_0 g$. We introduce dimensionless variables with the following measurement units: the mantle thickness D for length, D/k for velocity, D^2/k for time, T_0 for temperature, μ_0 for viscosity, $\eta_0 \ k/D^2$ for pressure and stresses, and kT_0/D^2 for heat source density.

[80] The convection equations (1–3) in these variables take the form

$$0 = -\partial p/\partial x_i + \partial S_{ij}/\partial x_j + \mathrm{Ra}T\delta_{i3} \ , \quad (5)$$

$$\partial T/\partial t + \mathbf{V}\nabla T = \partial(\partial T/\partial x_i)\partial x_i + H \ , \quad (6)$$

$$\partial V_i/\partial x_i = 0 \ , \quad (7)$$

where Ra is the Rayleigh number,

$$\mathrm{Ra} = \alpha\rho_0 gT_0 D^3/k\eta_0 \ . \quad (8)$$

Appendix A2: Equations of Motion of a Freely Floating Continent

[81] The Euler equations describing the horizontal motion of a solid continent of an arbitrary shape and its rotation around an instantaneous vertical axis reduce to a system of three equations [*Trubitsyn, 2000*]:

$$Mdu_0/dt = \iint(-p\delta_{1j} + S_{1j})n_j df \ , \quad (9)$$

$$Mdv_0/dt = \iint(-p\delta_{2j} + S_{2j})n_j df \ , \quad (10)$$

$$I_{33}d\omega_3/dt = \iint \varepsilon_{ij3}(x_i - x_{0i})(-p\delta_{jk} + S_{jk})n_k df \ , \quad (11)$$

$$dx_c/dt = u_0, \ dy_c/dt = v_0, \ d\varphi/dt = \omega_3 \ , \quad (12)$$

where I_{ik} is the tensor of moments of inertia of a solid

$$I_{ik} = \int \rho_c[(x_1 - x_{10})^2\delta_{lj} - (x_i - x_{i0})(x_k - x_{k0})]dV \ ,$$

$x_c(t)$ and $y_c(t)$ are the coordinates of the center of gravity of the continent, and φ is the rotation angle of the continent.

[82] In the particular case of horizontal motion of solid and infinitely thin continents, the viscous force is applied only to the base of the continent, for which we have $n_k = (0, 0, -1)$. Then the Euler equations are simplified and take the form

$$M\partial u_0/\partial t = -\iint S_{xz} dx \, dy \ , \quad (13)$$

$$M\partial v_0/\partial t = -\iint S_{yz} dx \, dy \ , \quad (14)$$

$$I\partial\omega/\partial t = \iint \left[(y' - y_0')S_{xz} - (x' - x_0')S_{yz}\right] dx'dy' \ , \quad (15)$$

where I is the moment of inertia of the continent about the axis z' passing through its center of gravity. Its value is calculated in a moving frame of reference with the axes x' and y' directed along the principal axes of inertia of the continent:

$$I = \int \rho_c \left[(x - x_0)^2 + (y - y_0)^2\right] d\tau \ . \quad (16)$$

[83] Dimension relations indicate that, as well as in the case of equation (1) describing the momentum transfer in a viscous fluid, the inertial terms on the left-hand side of Equations (9–11) are on the order of $k\rho/\mu \approx 10^{-23}$.

[84] Neglecting the inertial terms, the Euler equations yield six relations to find six unknowns: the coordinates of the center of gravity of the continent $x_c(t)$ and $y_c(t)$, its rotation angle φ and velocities $u_0(t), v_0(t)$, and $\omega_3(t)$:

$$\iint (-p\delta_{1j} + S_{1j})n_j df = 0 \ , \quad (17)$$

$$\iint (-p\delta_{2j} + S_{2j})n_j df = 0 \ , \quad (18)$$

$$\iint \varepsilon_{ij3}(x_i - x_{0i})(-p\delta_{jk} + S_{jk})n_k df = 0 \ , \quad (19)$$

$$dx_c/dt = u_0, \ dy_c/dt = v_0, \ d\varphi/dt = \omega_3 \ . \quad (20)$$

[85] The equation for the temperature (T_c) distribution inside a solid continent in the initial fixed reference frame reduces to the equation of thermal conduction with advective heat transfer:

$$\partial T_c/\partial t + \mathbf{u}\nabla T_c = \partial(\partial T_c/\partial x_i)\partial x_i + Q_c \ , \quad (21)$$

where Q_c is the density of heat sources inside the continent.

Appendix A3: Boundary Conditions

[86] The equations describing mantle convection (1–3), motion of a continent (9–11), and heat transfer in the continent (21) are interrelated though boundary conditions.

[87] The impermeability and free-slip conditions are set for mantle flows at the lower and lateral boundaries of the model region (the normal component of the fluid velocity and tangential components of viscous forces vanish at these boundaries):

$$V_k n_k = 0, \ S_{ki}\tau_i = 0 \ \ (i = 1, 2) \ , \quad (22)$$

where $n_k \ \tau_i$ are, respectively, unit vectors normal and tangential to a given surface.

[88] The impermeability and no-slip conditions are set at the boundaries of moving continents (velocities of the fluid mantle are equal to those of the continents):

$$V_i = u_i \quad (23)$$

at continental surfaces contacting with the mantle.

[89] The temperature at the lower boundary of the region is fixed, $T = 1$. The zero heat flux condition is accepted at lateral boundaries:

$$\partial T/\partial n_k = 0 \ , \quad (24)$$

where n_k is a unit vector normal to the boundary. At the upper free surface, the temperature vanishes ($T = 0$) only in the oceanic part, outside the continents. The continuity condition is accepted for the temperature and heat flux at the mantle-continent contact:

$$T = T_c, \ \partial T/\partial n = \partial T_c/\partial n \ . \quad (25)$$

The temperature is set to zero at the upper surface of a continent:

$$T_c = 0 \ . \quad (26)$$

[90] Thus, the mathematical problem can be summarized as follows. There three unknown functions of coordinates and time describing mantle convection: the mantle flow velocity vector $V_i(x, y, z, t)$, the temperature distribution $T(x, y, z, t)$, and the pressure distribution $p(x, y, z, t)$; we also have four unknown functions of time describing the motion of a continent as a whole: two components of the instantaneous velocity of the center of gravity $u_0(t)$ and $v_0(t)$, the instantaneous angular velocity of the rotation of a continent around its center of gravity $\omega(t)$, and the temperature distribution in the continent $T_c(x, y, z, t)$. These unknowns should be found by solving a system of interrelated equations: the differential equations of convection (1–3), three integral equations (17–19) resulting from the Euler equations, and the equation for heat transfer in the continent (21). Knowing at a given time moment the position of the continent and its velocities $u_0(t), v_0(t)$ and $\omega(t)$, its position at the next time moment can be found from (20). Boundary conditions (22–26) are used to determine constants of integration of the differential equations.

[91] The above problem with a freely floating continent differs from the known problem with a fixed continent by that the conditions for flow velocities and temperature at the boundary with a continent are set for the instantaneous position of a floating continent whose velocity and position are not known a priori but are determined by solving a system of interrelated differential equations.

[92] In the case of a few continents, the equations of motion (17–19) and the equation for temperature (21) are written out for each continent. In addition, the mutual impermeability condition is set for colliding continents. To this end, the forces of viscous friction of the continents are supplemented at the time of a local contact of the continents by the force pushing them away from each other; this additional force is applied at the contact area of the continents and is directed opposite to their relative velocity.

References

Anderson, D. L. (1989), *Theory of the Earth,* 366 pp., Blackwell, Boston.

Bobrov, A. M., W. Jacoby, and V. P. Trubitsyn (1999), Effects of Rayleigh number, length and thickness of continent on time of mantle flow reversal, *J. Geodynamics,, 27*(1), 133.

Fowler, C. M. R. (1996), *The Solid Earth,* 465 pp., Cambridge Univ., U.K.

Khain, V. E. (1994), *Basic Problems of Modern Geology,* 150 pp., Nauka, Moscow.

Schubert, G., D. L. Turcotte, and P. Olson (2001), *Mantle Convection in the Earth and Planets,* 941 pp., Cambridge Univ., U.K.

Scotese, C. R., L. M. Gahagan, and R. L. Larson (1988), Plate tectonic reconstruction of the Cretaceous and Cenozoic ocean basins, *Tectonophysics, 155,* 27.

Sepkoski, J. J., Jr. (1987), Periodicity in extinction and problem of catastrophism in the history of life, *J. Geol. Soc., 144,* 7.

Trubitsyn, V. P. (2000), Principles of the tectonics of floating continents, *Izv. Phys. Solid Earth* (in Russian), *36,* 101.

Trubitsyn, V. P. (2004), Evolutionary models of floating continents, *Russian J. Earth Sci., 6*(5), 311. (online version http//rjes.wdcb.ru/v06/tje04147/tje04147.htm)

Trubitsyn, V. P., and W. D. Mooney (2002), Mantle Convection with Thermally and Mechanically Coupled Floating Continents, *Open File Report 02–156,* p. 98, Geol. Surv., U.S.

Trubitsyn, V. P., and V. V. Rykov (1995), A 3-D numerical model of the Wilson cycle, *J. Geodynamics* (in Russian), *20,* 63.

Trubitsyn, V. P., and V. V. Rykov (2000), 3-D spherical Models of Mantle Convection with Floating Continents, *Open File Report 00–218,* p. 88, Geol. Surv., U.S.

Trubitsyn, V. P., and V. V. Rykov (2001), A numerical evolutionary model of interacting continents floating on a spherical Earth, *Russian J. Earth Sci., 3*(2), 83. (online version http//rjes.wdcb.ru/v03/tje01057/tje01057.htm)

Trubitsyn, V. P., W. D. Mooney, and D. A. Abbott (2003), Cool cratons and thermal blankets: How continents affect mantle convection, *Int. Geol. Rev., 45*(6), 479.

Turcotte, D. L., and G. Schubert (1982), *Geodynamics: Applications of Continuum Physics in Geological Problems,* 450 pp., John Wiley, New York.

Veil, R. M., R. M. Mitchem, and S. Tompson (1977), Global Cycles of the Relative Variations of the Sea Level, in *Seismic Stratigraphy – Application to Hydrocarbon Exploration, edited by Ch. Payton,* p. 25, Am. Assoc. Petrol. Geologists, Tulsa.

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