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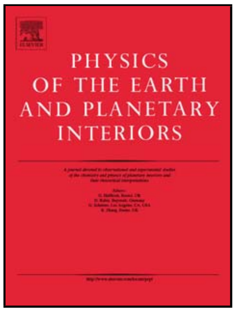
To cite this version:

Valeriy Trubitsyn, Mikhail K. Kaban, Marcus Rothacher. Mechanical and thermal effects of floating continents on the global mantle convection. Physics of the Earth and Planetary Interiors, 2008, 171 (1-4), pp.313. 10.1016/j.pepi.2008.03.011 . hal-00532142



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PII: S0031-9201(08)00052-6
DOI: doi:10.1016/j.pepi.2008.03.011
Reference: PEPI 4913

To appear in: *Physics of the Earth and Planetary Interiors*

Received date: 30-10-2007
Revised date: 12-3-2008
Accepted date: 14-3-2008

Please cite this article as: Trubitsyn, V., Kaban, M.K., Rothacher, M., Mechanical and thermal effects of floating continents on the global mantle convection, *Physics of the Earth and Planetary Interiors* (2007), doi:10.1016/j.pepi.2008.03.011

Mechanical and thermal effects of floating continents on the global mantle convection.

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Abstract

Numerical models are presented that simulate mantle convection coupled with superimposed continents. The continents and main islands are modelled as thin rigid spherical caps with non-slip boundary conditions and continuity conditions for temperature and heat flux at their bottom. Additional repulsive forces prevent overlap of the continents in the case of their collision. The initial temperature distribution in the mantle is calculated based on seismic tomography data. The evolving mantle model implies 10% basal and 90% internal heating and uniform viscosity. Mechanical coupling leads to near horizontal convection currents under continents and consequently to a noticeable decrease of the mantle temperature under them in spite of the thermal blanket effect. The modelling results show that a long-term evolution of the free convection model and of the model with implemented continents leads to principally different structures. Several common stages of the continental evolution are revealed. Back-arc basins at the active continental margins are closed at the first stage. In the next stage of the modelled evolution the convection pattern is reorganized and the main downwellings start to move to the south pulling the continents, which tend to assemble a new super-continent around Antarctica. Due to the repulsive forces the continents rotate and adjust to each other.

Key words: mantle convection, plate velocities, continental evolution.

Introduction

Paleomagnetic data reveal significant displacements of the continents with their periodic assemblage into a single super-continent with subsequent dispersal. The well-known super-continent Pangea existed about 250 Ma ago [Muller et al*.,* 1997], Rodinia - 750Ma [Torsvik et. al., 1995; Dalziel, 1994; Dalziel et al., 2000] and possibly Columbia – 1.8Ga ago [Zhao et al., 2002]. Formation and dispersal of the super-continents have been accompanied by opening and closing of the world oceans. The problem of organization of the global tectonics has been discussed for decades. However, up to now all theories are more or less phenomenologically based and are unable to explain quantitatively an underlying mechanism of many tectonics processes.

Latest studies have shown that besides internal instability of the global mantle convection, self-organization of the non-linear system representing continents coupled with the convecting mantle could be a reason for the global reorganizations of convective planforms (e.g. Gurnis, 1988; Trubitsyn and Rykov, 1995; Lowman and Jarvis, 1996; Bobrov et al., 1999; Lowman et al., 2001; Gurnis and Zhong, 2001; Trubitsyn et al., 2003; Phillips and Bunge, 2005, 2007; Trubitsyn et al., 2006). According to this hypothesis continents do not simply follow mantle flows but in return principally modify mantle convection patterns.

Considering a long-term evolution of the Earth one should differentiate the effects of the continents and large oceanic plates. At a time scale of less than ~100 Ma continents can be treated as frozen into oceanic plates. Therefore, plate tectonics works as a good approximation for the understanding of the global tectonics processes at this time scale. However, on the time scale of more than several hundred Ma the behaviour of the oceanic plates and continents is remarkably different: the oceanic plates are detached from the continents and continuously rejuvenated. Consequently, on the long-term time scale the end-member model of continents floating on the convective mantle with high near-surface viscosity could be sufficiently representative. Considering the intermediate time scale of 100 Ma - 1Ga one should use a more complex model with continents interacting with the oceanic plates and one needs to imply complex viscous-plastic rheology (e.g. Tackley, 2000; Zhong et al., 2000). However, due to the obvious complexity and computation problems, we have to study the end-member models at the moment.

Gurnis (1988) has created the first 2-D model that provides a quantitative description of the continents' aggregation and subsequent break-up. A continent was introduced as a high viscosity zone with an intrinsic chemical buoyancy at the top of the modelling box bounded by zones of low viscosity. This work was extended in further papers by incorporation of non-Newtonian rheology, tectonic plates, and faulted plate margins (e.g. Gurnis and Zhong, 1991; Lowman et al., 2001; Zhong and Gurnis, 1993, 1994, 1995; Lowman and Jarvis, 1996; Zhong et al., 2000). Boundary conditions for rigid plates floating on the convective mantle have been formulated by Gable et al. (1991): they imply force-balance conditions at the plate bondaries. These conditions together with the conditions for temperature and heat flux continuity for thick continents were applied for 2-D models to investigate evolution of the mantle-continent system in (Trubitsyn, 2000, Trubitsyn et al., 2003, 2006).

The thermal effect of the continents coupled with mantle convection was investigated in a series of papers (e.g. Cooper et al., 2004, 2006; Lee et al., 2005; Lenardic et al., 2005). The continents are defined by markers and the oceanic lithosphere is described by viscoplastis rheology (Cooper et al., 2004, 2006). It has been concluded that the nature of both, oceanic and continental thermal boundary layers is likely to be linked to preexisting strong chemical boundary layers. By contrast to a common view that continents, acting as thermal insulation above the convecting mantle, inhibit the Earth's internal heat loss, the authors have demonstrated that increased continental insulation might have no effect on mantle heat flow and can even enhance it. In general, the local thermal structure of oceanic and/or continental lithosphere cannot be modeled independently of considering the global ocean-continent system.

A 3-D Cartesian model with incorporated rigid continents was investigated by Trubitsyn and Rykov (1995). Velocities of the continents and mantle flow were calculated selfconsistently by solving a system of coupled equations. The equations for rigid continents describe force and torque balance. For a 3-D spherical case a complete system of the equations, implying thermal and mechanical coupling of the thick continents and convective mantle, has been suggested by Trubitsyn and Rykov (2000, 2001). Preliminary spherical models with several floating continents were investigated in the abovementioned papers. These models already imply thermal and mechanical coupling of the mantle and continents as well as the effects of continental collision and rotation. However the modelling grid was very coarse and the Rayleigh number very low, conditions that do not permit to reproduce the system sufficient accuracy. Phillips and Bunge, (2005) have developed very fine convection models with a single ring-shaped continent using the FE code Terra. Mechanical effects of the coupled system have been modelled implying non-slip boundary conditions and thermal effect – by fixed temperature at the bottom of a thin spherical cap floating on the mantle. Phillips and Bunge (2007) have considered six continents, which are approximated as buoyant, perfectly isolated spherical caps. They did not calculate the rotation of the continents because of their round shape. They also did not incorporate repulsion forces because the continents did not collide, being independent throughout the model evolution. These authors have found that a regular super-continental cycle, which occurs in the idealized model with three continents floating on the internally heated mantle, lasts for about 400my. Periodical assemblage and disperse of 12 colliding continents (round shape) were also demonstrated by Trubitsyn and Rykov (2001), however with the model having a low Rayleigh number and a low resolution.

In the present study we continue this work and analyse a spherical model of the convecting mantle with 15 floating bodies representing real continents and major islands. We study principal effects of mechanical and thermal coupling of the continents and mantle. Comparing to the early model of Phillips and Bunge (2005, 2007) we introduce two features to the global model. First, instead of rounded continent**s** we consider the full system of the existing continents and main islands. Second, the continents are not just mechanically coupled with the convecting mantle but can also interact with each other. The main goal is to demonstrate that this system is self-organized and the major stages of the global long-term tectonics do not appear chaotically but are intrinsic characteristics of the dynamic Earth.

This work continues the study of the long-term evolution of the mantle-continents system (Trubitsyn et al., 2006). The previous 2D model of a thick continent floating on the convecting mantle has already revealed several principal stages of evolution of the active continental margins: a) Extension tectonics and marginal basins form at the boundary of a continent approaching to the subduction zone, roll back of subduction takes place in front of moving continent; b) The continent reaches the subduction zone, the extension regime at the continental edge is replaced by strong compression. The roll back of the subduction zone still continues after closure of the marginal basin and the continent moves towards the upwelling (Trubitsyn et al., 2006). As a result the Pacific Ocean becomes non-symmetric; c) The continent overrides the upwelling and subduction in its classical form stops. In the present paper we start to test these conclusions on 3-D spherical models.

Numerical model

"Free" thermal convection is simulated in a 3-D spherical shell with free-slip, zero radial velocity and constant temperature conditions at the outer and inner surfaces. The possible proportion of the internal and basal heating has been discussed for decades. It was proposed that the basal heating does not exceed 8% of the total amount (e.g. Davies, 1999, Schubert et al., 2001). However, Bunge (2005) has recently shown that the radial distribution of superadiabatic temperature better fits to a higher amount of the basal heating (up to 25%). Due to chemical layering (e.g. Tackley, 2002) most of the heat sources might be concentrated in the lowermost mantle, while the mantle is depleted.

The dimensionless equations of the mantle convection (see appendixA1-A5) contain two principal parameters: Rayleigh number Ra_R and thermometric density of heat sources H. We take them equal to $Ra_R=1\cdot 10^7$ and H=100. Compared to the parameters of the real Earth (see Appendix) the overall mantle viscosity of the model is ~18 times greater while the internal heating is ~1.4 times less than the real one. As a result the effective Rayleigh number is approximately 25 times less, which leads to lower mantle velocities (about 10 times compared to the real Earth). Dimensionless bottom and surface heat fluxes, which correspond to these parameters, are $q_c=10$ and $q_s=30$, respectively. The low Rayleigh number provides the possibility to perform calculations on a relatively dense grid (58x144x216, $\Delta r=50$km, $r\Delta\theta=140$km and $r\Delta\varphi_{max}=200$km) with sufficient accuracy (about four points in the thermal boundary layer).

Therefore, similarly to Phillips and Bunge (2005, 2007) we still employ the models, which are different from the real Earth, however provide the possibility to investigate specific effects related to the interaction between the continents and global mantle flow. Though our numerical code provides the possibility to take into account both, radial and lateral viscosity variations, it requires enormous computer time for the convergence of the iterative process, which is used for these purposes. For that reason at the initial stage we test the model with constant mantle viscosity and a low Rayleigh number. Since the effective Rayleigh number for the real mantle is ~25 times higher, the obtained results (flow and plate velocities) are properly scaled as proposed in (Schubert et al., 2001, Phillips and Bunge, 2007). The governing equations and a detail description of the numerical model can be found in the Appendix. Here we discuss only some principal aspects. The lateral temperature variations within the Earth are based on the existing tomography models. The initial radial temperature profile shown in Fig. 1 is obtained from a preliminary model of "free" convection without continents after achieving a steady state with the above parameters (the Rayleigh number and the proportion of basal and internal heating). The continents are modelled as thin rigid caps, the initial position and shape of which are taken from geographical maps. The caps are mechanically coupled with the convecting mantle (non-slip conditions at the bottom of the caps). For more details see Appendix.

To investigate the thermal blanket effect we introduce appropriate boundary conditions for the continents. Phillips and Bunge (2005, 2007) have previously considered simplified models, in which some fixed temperature or perfect isolation are set at the base of the continent. However, during evolution actual temperatures and heat flux may significantly vary. In the present model we employ more realistic continuity conditions for temperature and heat flow at the base of a continent. According to the Eqs. A12, the temperature T(h) inside the continent with heat production $H_c = H_{cs}$ $\exp(-h/h_r)$ may be determined as a function of depth as follows (Turcotte and Schubert, 1982):

$$T(h)=T_s+q_m h/k_c+(H_{cs}h_r^{2}/k_c)[1-\exp(-h/h_r)], \tag{1}$$

where T_s is the surface temperature, q_m - the mantle heat flow, k_c - the thermal conductivity, H_{cs} - the effective surface heat production**,** h - the depth and h_r - the length scale. Therefore, the heat flow at the base of the continent is equal to:

$$q_m= (k_c/d)(T- T_s -T_r), \tag{2}$$

where d is the thickness of the plate ($d >> h_r$), T_r is the temperature excess due to the heat production $T_r= H_{cs}h_r^2/k_c$. The heat flow in the mantle is $q_m=-kdT/dr$ and the thermal boundary condition at the base of the continent is simply:

$$-dT/dr=\xi(T-T_s-T_r), \tag{3}$$

where ξ characterizes an effective rate of the thermal blanket effect of the continent. In a dimensionless form $\xi=(R/d)(\kappa_c/\kappa)$. If $\xi =0$, the continent fully prevents heat escape from the mantle (100% blanket effect), while, if ξ is infinity, the blanket effect is neglected. These conditions might then be used without considering a specific thickness of the continents.

In the Appendix one can find more details of the numerical model. The numerical code is the same as in (Trubitsyn and Rykov, 2000, 2001), however it was adopted for a much higher Rayleigh number and resolution. We tested the numerical code by comparing the results with other available programs. Without continents our results are in agreement with the results obtained for the same models developed with the code CITCOM (Zhong et al., 2000; Tan et al., 2006). We have also found a qualitative correspondence with the results obtained by Phillips and Bunge, (2005) with the code TERRA. Benchmarks for the floating continents coupled with the convecting mantle do not exist. Therefore, to check the stability of our results we performed calculations for a wide range of the initial parameters: density and size of the grid, various values of the Rayleigh number and various parameters of continents, their number, shape and size. In all cases the obtained results show qualitative similarities and predicted tendencies. In the following chapters we discuss only the effects, which appear in all models.

In the present work we have performed calculations on a grid 58x144x216 ($r/\theta/\varphi$). Note that the resolution of this grid is even higher than the resolution of the existing global tomography models.

RESULTS

1. Free steady state of mantle convection

As already mentioned, in the first step we performed calculations for the free convection model with basal and additional internal heating ($Ra_R=10^7$ and H=100). The consistency of the results was tested for different grids (from 58x36x72 to 58x144x216) and various initial conditions after the convection achieves quasi steady state. The calculated mantle convection pattern is similar to the results obtained for the model with internal heating in (Bunge and Richards, 1996 and Phillips and Bunge, 2005, 2007).

The results for free steady state convection are shown in Fig. 2. The calculated dimensionless temperature and mantle velocities are shown for a depth of 100km. The basic grid is 58x144x216. We also demonstrate the dimensionless velocities and lateral temperature variations relative to the mean temperature for a sample cross-section ($\varphi=0$). The calculated radial superadiabatic temperature distribution T(r) and radial heat flux q(r) are shown in Fig. 1.

It is clear that the calculated thermal boundary layer is about 150km (Fig. 1), which correspond to three-four grid points. The calculated surface heat flux q_s is about three times higher than that at the bottom (q_c). This means that the basal heating $Q_c=4\pi r_c^2 q_c$ is approximately 10 times less than the total surface heat flux $Q_c=4\pi r_c^2 q_c$ (the basal and internal heating are related as 10% and 90%).

The calculated dimensionless mean velocity is equal to $v_m= 1200$. When properly scaled (see Appendix) it corresponds to the mean velocity of about 0.6 cm/y. Keeping in mind that the mean mantle velocity of the real Earth is about 5cm/y, the calculated velocities should be increased by a factor of 10 and the time intervals should be correspondingly decreased. The calculated thermal boundary layer is about 3 times thicker compared to the real Earth (~150 km, Fig. 1). Therefore, the Nusselt number is approximately 3 times less (the effective Rayleigh number is about 30 times less, $Nu\sim Ra^{1/3}$) compared to the real Earth. These estimate agree with the empirical relationship for velocities ($v\sim Ra^{2/3}$).

2. Mantle convection with moving continents. Initial conditions

As initial radial temperature distribution we take the calculated superadiabatic temperature profile for the quasi-steady state model, while lateral temperature variations are based on the S20a seismic tomography model (Ekstrom and Dziewonski, 1998). The V_s perturbations have been initially converted into density variations applying a constant velocity-to-density scaling factor $d\ln\rho/d\ln v_s=0.25$ (e.g. Karato, 1993). The density perturbations were converted into temperature variations ($d\rho=-\alpha\rho dT$) applying the depth-dependent thermal expansion coefficient (Paulson et al., 2005):

$$\alpha = \left(3 - \frac{2h}{h_{cm}}\right)10^{-5}, \qquad (4)$$

where h is the depth and h_{cm} the depth to the core-mantle boundary. It is clear that not all velocity variations presented in the S20a model are induced by temperature variations. Compositional variations could substantially alter the velocity-temperature relationship or simply mask the temperature effect. However, we assume that in spite of possible alterations of the computed models, general conclusions about the importance of incorporating floating continents in global dynamic models will be convincing.

With this initial temperature distribution we solve the equations of the mantle convection coupled with floating continents, the positions of which are similar to the present day Earth. As explained above, for the mechanical coupling we apply non-slip boundary conditions at the bottom of the continents. The thermal coupling is taken into account by continuity of temperature and heat flow at the continent bottom as described above (Eq. 3). Taking a continent thickness (including lithosphere) d=300km and a thermal conductivity $\kappa_c/\kappa=0.6$, we arrive at the effective blanket ration $\xi=(R/d)(\kappa_c/\kappa)=13$.

The calculated temperature at a depth of 100km and the continental/mantle velocities (after the initial time step) are shown in Fig. 3 together with the velocities according to the REVEL model [Sella et al., 2002] based on GPS data. Similar to the free convection, the calculated mantle velocities are about 10 times less than for the real Earth. Note that the considered model does not take into account oceanic lithospheric plates. Nevertheless the calculated surface velocities sufficiently reasonably resemble the observed ones.

3. First stage of the evolution. Adjustment of the initial low-resolution tomography data to the finer model grid.

As soon as we start to calculate the time evolution of the model, the continents start to move and the convection pattern is changed. Note (e.g. Trubitsyn and Rykov, 2000, 2001), that mantle convection with moving continents never achieves steady state.

The calculated patterns of mantle convection coupled with floating continents are shown in Fig. 4. The resolution of the tomography model does not exceed 20th degree (approx 1000x1000km), which is more coarse than the numerical grid (max. 140x185 km). Therefore, at time t=0 the upwellings and downwelligns are very broad. However, in a short time (tens of Ma) the convection pattern becomes much finer, adjusting to the calculating grid resolution. During the initial time period we mostly observe reorganization of the mantle flow pattern but the continents do not move noticeably.

4. Long-term drift of the continents floating on the convective mantle

The main steps of the continental drift for different time intervals are shown in Fig. 5. The continental velocities are always 3-5 times less than the maximal velocities of the mantle flow beneath them. For instance, Africa has moved to the west for about 900 km during a time period of 100 Ma, which corresponds to the horizontal velocity ~1 cm/y. Therefore, floating continents coupled with the mantle stabilize the evolution of the convection pattern. A general tendency could be gained from the comparison of the computed models: the continents and islands always have a tendency to assemble into larger groups. Comparing the initial positions of the continents in Fig. 5 (t=0) with the positions at t=10My and 40My one can see that Eurasia is permanently moving to the South-East. As a result, the distance between Eurasia and Japan reduces and the Sea of Japan is closed after approx. 20My. This result agrees with the results based on the 2D modelling (Trubitsyn et al., 2006), which predicts the closing of marginal basins at the second stage of the evolution of the active continental boundaries and the replacement of an extension regime by strong compression.

The further movement of Eurasia in this direction leads to a decrease in the distance between Eurasia and Australia and an assembly of the islands located between these continents. As a result, at time 50-100 Ma Africa, Eurasia and Australia start to form a new super-continent. At the same time, South America approaches Antarctica. At the time of about 20 Ma Madagascar begins to move away from Africa towards Eurasia and approaches India at about 60Ma. Starting from this time Eurasia closes the gap with Australia and pushes the islands (Borneo and Sumatra) aside. And finally, at the time of about **1**00Ma, Africa, Eurasia, Australia, Antarctica and South America form a new group (future super-continent). At the same time North America moves slowly to the south following South America. In general all continents tend to move from the northern to the southern hemisphere**.**

When comparing different stages of the continental evolution it is clear that there are two main factors controlling this process. First, these are massif subduction zones, mostly on the eastern and southeastern borders of Eurasia, which pull the continents in the southeastern direction. Second, these are upwellings associated with Southwest Africa and North Atlantic including the Polar Ocean permanently pushing Africa and Eurasia correspondingly.

5. Effects of thermal and mechanical coupling of the mantle and continents on a long-term evolution.

One of the final evolution stages, calculated for the model, is shown in Figs. 6 and 7. In the case of only mechanical coupling, the mantle under the continents is remarkably colder. During the whole evolution time the convective flows under the continents remain horizontal, which reduces the advective vertical heat transfer. The thermal blanket effect produces an opposite effect, however it takes a much longer period of time to accumulate enough heat under continents to see a substantial difference to the model without thermal resistance. The thermal blanket effect is especially visible when the continent velocity is low while the horizontal size is large. This result is in agreement with the results of 2D modelling (Bobrov et al., 1999) demonstrating that the time for a heat accumulation is about 300Ma for a normal-size motionless continent and about 100 Ma – for a super-continent.

Due to viscous drag, the continents are shifted towards downwellings and at time 100 My occupy most of them, which is clearly visible in Figs. 6 and 7. Comparing the mantle convection pattern with floating continents (Fig. 7 A,B) and the quasi-steady state free convection with the same parameters and initial conditions (Fig. 7 C), one can see that continents have principally changed the face of the Earth, while the convection pattern in Fig. 5 (free convection) is similar to the quasi-steady state (Fig. 2). We stopped the evolution of the model approximately at 100 Ma. A further development needs more advanced mechanism for the continents' interaction, more realistic mantle parameters and large computing time. This is the task for the nearest future.

Discussion and conclusions

Mechanical and thermal effects of the floating and colliding continents on the convection patterns with applications for the tectonic history of the Earth have been previously studied for 2D models (Trubitsyn et al., 2006). In the present paper we try to understand if principal results of the 2D modelling are also applicable to a 3-D spherical model. Importantly, in our 3-D model we provide detailed calculations including interaction between colliding continents.

For the initial stage we do not take generally accepted ideal steady state convection. On the other hand, for the first try we used the initial conditions, which are close to the Earth structure as it is reflected in the tomography model. It is clear that the model is still far from the real Earth, nevertheless it provides a possibility to study such effects as continental rotation and collision.

The numerical evaluation of 3-D spherical convection models with implemented continents still needs enormous computer time (at least several weeks with a supercomputer) as in (Phillips and Bunge, 2005) and this study. These limitations enforce us to simplify the analysed models, e.g. to use the reduced calculation grids and a lower Rayleigh number, i.e. to increase mantle viscosity compared to the real Earth. As well we only consider a relatively short period of the evolution (up to 100My). Therefore these results should be considered as preliminary, however we are permanently moving towards the real planet. Comparing the results obtained for simplified models with those computed for high-resolution grids and greater Rayleigh numbers (for short time intervals), we may conclude that the results are qualitatively very similar if not considering small-scale details, which are beyond this study. This gives us every reason to consider the obtained results as general with certain limitations.

Also due to technical difficulties we still do not consider oceanic plates and the effects related to non-zero continent thickness. The thermal effects of thick continents on long–time mantle convection were investigated in details on 2-D models in several papers. Cooper et al. (2004; 2006), Lee et al., (2005) and Lenardic et al., (2005) have demonstrated that continental insulation does not generally reduce global heat flow. Such insulation can have a negligible effect or even enhance mantle cooling. Their theory also suggests a potential constraint on the continental surface area.

Increased surface area enhances the subduction rate of oceanic lithosphere. The authors found that the mantle heating due to the thermal blanket effect is more effective than its cooling due to a redistribution of radioactive elements from the mantle into the continental lithosphere. On a large time scale, the thermal blanket effect raises mantle temperature, decreases viscosity and consequently increases the Rayleigh and Nusselt numbers. This allows for more rapid overturn of oceanic lithosphere, which can, depending on continental area, increase global heat loss (Lenardic et al., 2005; Cooper et al., 2006).

Lenardic and Moresi (1999) have concluded that the buoyancy caused by chemical depletion is also unlikely to stabilize cratonic roots alone, the roots must be 10^3 times as viscous as the mantle. On the other hand, Trubitsyn (2000) and Trubitsyn et al. (2003, 2006) have demonstrated that the thermal blanket might produce principally different effects for the motionless and floating continents as well as for the small and large ones. Warming of the mantle under the motionless continent (normal size) needs significant time (more than 300My). During this time period the floating continent could pass a distance of about 3000 km (1 cm/y) and therefore move away from the warmed place. As a result, the thermal blanket can not significantly change thermal state of the mantle under even slowly moving continent.

Assemblage of continents on a large time scale was studied by evaluating several numerical models in (Trubitsyn, 2000, 2006). It was shown that after assemblage of two nearest continents, the increased downwelling may appear under them. This downwelling can pull other continents located at some distance from the coupled ones. As a result most of the continents assemble in a unified supercontinent above a large downwelling. However, insufficient resolution and a reduced Rayleigh number don't permit to investigate fine structure of these processes.

The thermal blanket effect, which is demonstrated by Cooper et al. (2004; 2006), Lee et al., (2005) and Lenardic et al., (2005), plays an important role only for large and slowly moving continents (e.g present Africa or supercontinent Pangea). After formation, a supercontinent remains unmoving for sufficiently long time period, which leads to mantle warming and in the end to reversal of the mantle currents with a subsequent break-up, which is also confimed by (Trubitsyn et al., 2003). Hot upwellings, generated by the thermal blanket under a slowly moving supercontinent, can lift it to about 100m. It was shown (Trubitsyn, 2000) that **s**uch vertical uplifts of the continents might correlate with the observed Phanerozoic global sea changes (Vale et al., 1977) and with the global extinction of plants and animals (Sepkoski, 1987). Some correlation could be traced with duration of the Paleozoic, Mesozoic and Cenozoic eras. Therefore, the difference of the results obtained in the abovementioned studies with the present one may be explained by the different time-scales of the considered processes.

The present study confirms principal results obtained for the low resolution models with lower Rayleigh number, however it reveals additional features of the processes of mantle currents reorganization. It is shown in Figs. 6 and 7 that the feed-back of the moving continents principally change the mantle convection pattern. One of the new effects is concentration of the downwellings under a group of assembling continents already at time of about 100My.

The main conclusions of our analysis are follows:

1. Drag-pull forces move floating continents toward the nearest downwellings. As a result, the continents approach the coldest zones in the upper mantle. In turn, the continents modify the mantle structure since they are moving coupled with downwellings. This effect appears already at the short time-scale of about several tens Ma, since mantle velocities nearly instantaneously adjust to the mantle structure. As a result, downwellings tend to concentrate under a group of continents. The thermal blanket produces the opposite effect but needs a substantially longer time period.

2. The inclusion of the repulsive forces gives us a possibility to model rearrangements of the continent system. This model allows us to calculate not only displacements of the continents, but also the formation of a super-continent. The continents’ assemblage is accompanied by a rotation of the continents, which provides a compact package and induces an additional toroidal component of the plate velocities.

3. With the initial structure corresponding to the present state of the Earth (as derived from seismic tomography data) the models with implemented continents reveal a tendency to close marginal seas and to shift subduction zones under the continental margins. After an evolution of about **1**00Ma, the continents start to move to the southern hemisphere with a tendency to form a new super-continent.

4. The 4-D modelling allows us to understand if the initial-state parameters are stable with respect to a further evolution of the mantle convection, in other words, if the dynamic model is self-consistent. This gives additional constraints for the construction of the present-day (snap-shot) model of the dynamic Earth. This model should not only fit to the observed parameters but should also be stable with respect to a further evolution, which corresponds to the general tendencies revealed in the Earth history. The real mantle structures should satisfy the equations of momentum,

mass and heat transfer. Only the modelling of the evolution provides a possibility to check all these conditions. We demonstrate that the temperature distribution derived from the seismic tomography data changes with time. The reason for these changes could be roughness of the tomography data, not correct transformation to the temperatures, or most probably - neglecting lithospheric plates and some additional effects and parameters of the real Earth in the analysed models (e.g. thickness of the continents and 3-D viscosity distribution).

5. The developed dynamic models are preliminary; however, these results reveal many features, which are found from geological reconstructions. We do not pretend to consider our results as any prediction of the future evolution of the Earth. We only try to demonstrate, that the model of the mantle convection with floating continents is self-organized in comparison with the stochastic “free” convection model.

6. Despite many simplifications of the analysed models, the obtained results demonstrate the principal possibility to construct, in the nearest future, a self-consistent model of the convecting mantle coupled with floating lithospheric plates (continental and oceanic).

Acknowledgements.

We are grateful to V. Rykov for his important contribution in developing the numerical codes used in the study. We would like to thank H.-P. Bunge, P.J. Tackley and M. Jellinek for very constructive advices and for stimulation to make additional calculations for more realistic models of the mant1e convection. This work was partially supported (Trubitsyn) by the Russian Foundation of Basic Research (grant 08-05-00378).

APPENDIX

Equations and boundary conditions for the mantle convection coupled the floating continents

The mantle convection coupled with floating continents is modelled using classical Stokes' equations for viscous flow and the equations describing the motion of a floating rigid body.

To make the paper more compact and to clarify the main equations and boundary conditions for the mantle convection with moving continents, we introduce them in a simple Cartesian frame. However, the basic equations can be rewritten for a spherical frame using available textbooks.

A-1. Equations of thermal mantle convection

Thermal convection in the mantle is described by distributions of the convective velocity vector $V_i(x,y,z)$, temperature $T(x,y,z)$ and pressure $p(x,y,z)$. These unknown functions can be found from a solution of the system of three equations describing the momentum, heat, and mass transfer. In a dimensionless form and assuming the Boussinesq approximation we have the following equations (e.g. Schubert et al., 2001):

$$0 = -\partial p/\partial x_i+\partial S_{i\,j}/\partial x_j+ Ra_{\mathbf{R}}\mathbf{T}\,\delta_{i3} \qquad \text{(A1)}$$

$$\partial T/\partial t+V\nabla T = \underline{\kappa}\partial(\partial T/\partial x_i\,)\partial x_i\ + H \qquad \text{(A2)}$$

$$\partial V_i/\partial x_i = 0 \qquad \text{(A3)}$$

where the viscous stress tensor is equal to:

$$S_{i\,j} = \eta\ (\partial V_i\,/\partial x_{j\,+}\,\partial V_j\,/\partial x_i) \qquad \text{(A4)}$$

and Ra_R is the Rayleigh number, scaled to the Earth's radius

$$Ra_R=\alpha\rho_0 gT_0R^3/(\kappa_0\eta_0\,) =(R/D)^3Ra_T \qquad \text{(A5)}$$

Here $H=H_m/c_p$ is the thermometric density of internal heat sources, H_m is the internal heating rate per mass unite, T is the superadiabatic temperature and $Ra_T = \alpha\rho_0 gT_0D^3/(\kappa_0\eta_0)$ is the Rayleigh number defined by a temperature difference. As units of the calculated parameters we take: the Earth's radius R for length, $V_0=\kappa/R$ for velocity, $t_0=R^2/\kappa$ for time, T_0 for temperature, η_0 for viscosity, c_{p0} for thermal capacity, κ_0 for the thermal diffusivity or $k_0 =\kappa_0(\rho_0c_{p0})$ for the thermal conductivity, $\eta_0\kappa/R^2$ for pressure and stresses, and $H_0=\kappa T_0/R^2$ for thermometric density of heat sources. Parameters for the real Earth usually are taken as follows (Schubert et al., 2001) D=2900km, R=6370km, $\alpha = 2.0\cdot 10\text{-}5K^{-1}$, $\kappa_0=10^{-6}$, $\rho_0=4.6\cdot 10^3$ kgm^{-3}, $c_{p0} = 1.1\cdot 10^3$ J/kgK, $\eta_0=3\cdot 10^{21}$, $k_0=\rho_0c_p\kappa_0=5W/(mK)$, $H_m=9.23\cdot 10^{-12}W/kg$, $H=8.4\cdot 10^{-15}K/s.$, $q=72mW/m^2$. Taking $T_0=2500K$ the units are equal to $t_0=R^2/\kappa=4.05\cdot 10^{19}s =1.3\cdot 10^{12}y$, $V_0= \kappa/R=R/t_0 =4.9\cdot 10^{-6}m/y$, $H_0= \kappa T_0/R^2=6\cdot 10^{-17}$ K/s. With these units we obtain for the dimensionless thermometric density of heat sources $H/H_0=140$, and for Rayleigh numbers $Ra_T=\alpha\rho_0 gT_0D^3/(\kappa_0\eta_0)=1.8\cdot 10^7$, $Ra_R=\alpha\rho_0 gT_0R^3/(\kappa_0\eta_0)=1.9\cdot 10^8$, $Ra_H=\alpha\rho^3{}_0g\ H_m\ c_{p0}D^{\,5}/(k^2\eta_0\,)=1.2\cdot\ 10^9$.

A-2. Equations for a floating continent

As first approximation, we consider continents as infinitely thin rigid bodies, the motion of which is described by Euler's solid body equations. The continent is driven by the forces originating from the mantle flow. Neglecting inertial forces and considering only the rotation of the continent around the vertical axis we can write six equations for the six unknown parameters: the coordinates of a centre of gravity $x_c(t)$, $y_c(t)$, rotation angle φ , and continent velocities $u_0(t)$, $v_0(t)$, $\omega_3(t)$:

$$\iint (-p\,\delta_{1j} + S_{1j})n_j\,df + \sum_n F^n_1 = 0, \tag{A6}$$

$$\iint (-p\,\delta_{2j} + S_{2j})n_j\,df + \sum_n F^n_2 = 0, \tag{A7}$$

$$\iint \varepsilon_{i\,j3}\,(x_i - x_{0i})(-p\,\delta_{jk} + S_{jk})\,n_k df + \sum_n F^n_i\,(x_i - x_{0i})\varepsilon_{i\,j3} = 0, \tag{A8}$$

$$dx_c/dt = u_0,\quad dy_c/dt = v_0,\ d\varphi/dt = \omega_3 \tag{A9}$$

where ε_{ijk} is the Levy-Civita symbol, which is equal to 0 if any two indices coincide, 1 for the even transposition of the indices with respect to (1,2,3), and -1 if the transposition is uneven. Through F^n_i we denote unknown forces, which act on a continent from neighbouring continents at each point of their contact. These forces are estimated by a “trial and error” iterative method to satisfy the basic condition: the continents should not overlap each other. The integrals are taken for the whole surface of a continent submerged into the mantle. The viscous tensor of mantle flow S_{ij} is a function of the mantle velocities (Eqs. A4).

In our particular case of a horizontal motion of the rigid thin continents, the viscous force acts only at the base of the continent where $n_k=(0,0,-1)$. Then, the Euler’s equations are simplified:

$$0 = \iint S_{xz}\,dx'\,dy' + \sum_{\mathbf{n}} \mathbf{F}^{\mathbf{n}}_{\mathbf{x}},\ 0 = \int S_{yz}\,dx'\,dy' + \sum_{\mathbf{n}} \mathbf{F}^{\mathbf{n}}_{\mathbf{y}},$$

$$0 = \iint [(y' - y'_0)S_{xz} - (x' - x'_0)S_{yz}]\,dx'\,dy' + \sum_{\mathbf{n}} \mathbf{F}^{\mathbf{n}}_{\mathbf{y}}\,(\mathbf{x_n} - \mathbf{x_0}) + \sum_{\mathbf{n}} \mathbf{F}^{\mathbf{n}}_{\mathbf{x}}\,(\mathbf{y_n} - \mathbf{y_0}) \tag{A10}$$

Velocity $\mathbf{u}$ of any point of the rigid continent is equal to:

$$\mathbf{u}(\mathbf{r}) = \mathbf{u}_0 + [\omega_3(\mathbf{r} - \mathbf{r}_0)]. \tag{A11}$$

Therefore, for the calculation of the continental velocities we need to calculate only the velocity of the centre of gravity $\mathbf{u}_0=(u_0, v_0)$ and the angle rotation velocity ω_3.

The equation for the temperature T_c inside the continent (in the original co-ordinate system) is the heat conduction equation, which includes the advective heat transfer resulting from continental movement (velocity from Eq. 9):

$$\partial T_c/\partial t + u\nabla T_c = \underline{\kappa_c}\partial(\partial T_c/\partial x_i)\partial x_i + H_c \tag{A12}$$

where κ_c and H_c are thermal diffusivity and thermometrical density of the heat sources within the continent.

A-3. Boundary Conditions

The equations of mantle convection (A1 - A3), continent motion (A6 - A9) and heat transfer (A12) are interconnected through boundary conditions.

As mentioned above, the impermeability and free-sip conditions are assumed for the mantle flows at the bottom and upper surface of the mantle outside the floating bodies; i.e. a normal component of the fluid velocity and tangential components of the viscous force are all equal to zero,

$$V_k n_k=0,\ \ S_{ki}\tau_i=0,\ i=1,2. \qquad \text{(A13)}$$

where n_k and τ_I are the unit vectors, normal and tangential to the surface respectively.

The impermeability and non-slip conditions are assumed at the bottom of the floating bodies, i.e. the mantle flow and continent velocities are equal:

$$V_i=u_i \qquad \text{(A14)}$$

The temperature at the bottom is fixed (T = 1 in a non-dimension case). T=0 at the upper free boundary of the mantle outside the continent. The temperature and heat flow continuity conditions are assumed at the mantle-continent interface,

$$T = T_c\ ,\ \partial T/\partial n_{\mathbf{k}} = \partial T_c/\partial n_{\mathbf{k}} \qquad \text{(A15)}$$

as explained in the main text (Eqs. 1-3).

Above in the text it was shown, that for a thin continent equation (A15) can be replaced by (3)

A-4. Numerical scheme

The mathematical problem can be summarized as follows. There are three unknown functions describing the mantle convection (the velocity vector of mantle flows $V_i(x,y,z,t)$, the temperature distribution $T(x,y,z,t)$, and pressure $p(x,y,z,t)$) and four unknown functions describing the motion of the rigid bodies (two velocity components for each centre of gravity $u_0(t)$ and $v_0(t)$, the angular velocity ω(t) for a rotating body and the temperature distribution within the continent $T_c(x,y,z,t)$. They can be found from the system of the following coupled equations: three differential equations of the convection (A1 - A3), three integral relations (A6 - A8) derived from the Euler's equations, and the equation of heat transfer in the continent (A12).

To solve the system of differential equations of mantle convection with boundary conditions we have to find the velocities **u** at each continental boundary, which are unknown and depend on u_0, v_0 and ω_3 (A11)**.** To overcome this uncertainty, we use a trial and error method for finding u_0, v_0 and ω_3. The convective flow velocities, calculated with these continental velocities as boundary conditions, should fit to the integrals (A6-A9) with some predefined misfit. If the position and velocity of the continent are known at a given moment, the position at the next time moment can be found from **(**A9**)**. The integration constants for the differential equations are defined from the boundary conditions (A13 - A14). Therefore we determine the boundary conditions at the continent-mantle interface (temperature and flow velocities) at each step by a solution of the connected equation systems, but not consider them as being fixed.

In the case of several continents, the equations of motion (A6 – A9) and temperature (A12) are solved for each continent. If two continents collide, additional impermeability conditions are formulated. For this purpose we introduce repulsive forces at any point where the continents touch each other, the forces are opposite to the continental velocities. Again, they are determined in an iterative procedure by trial and error. The values of the repulsive forces should fit the condition that the continents do not overlap and do not move away for more than some predefined distance that specifies the accuracy of the calculations. These repulsive forces can be used for the calculation of continents' deformations, but we do not apply this procedure in the present paper.

A finite difference method is used to solve the above given systems of equations (Trubitsyn and Rykov, 2001). The flux-corrected transport algorithm of Zalesak (1979) is applied to solve the heat conservation equation. The equations for the velocities and pressure, respectively, are reduced to elliptical equations with variable coefficients (generalized Poisson equations). They are solved using the three-layer modification of the triangular method and the method of conjugate gradients for chosen iteration parameters.

We have performed calculations for the model grid of 58x144x216 (r/θ/φ). Since the size of the large islands and microcontinents is already comparable with the size of one grid cell, we have used a special procedure to take into account the small-scale features similar to the commonly used markers. The calculated near-surface flow velocities were interpolated to the high-resolution hyper-grid (10 times finer than the basic one), which was used then to calculate displacements of the continents and islands. In fact, the continents are modelled as 2-D active markers, which may not only move on the surface like normal point-markers but can also rotate. At the same time, these 2-D continental markers actively interact with each other and with the surrounding mantle via the specified boundary conditions and repulsive forces.

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Figures

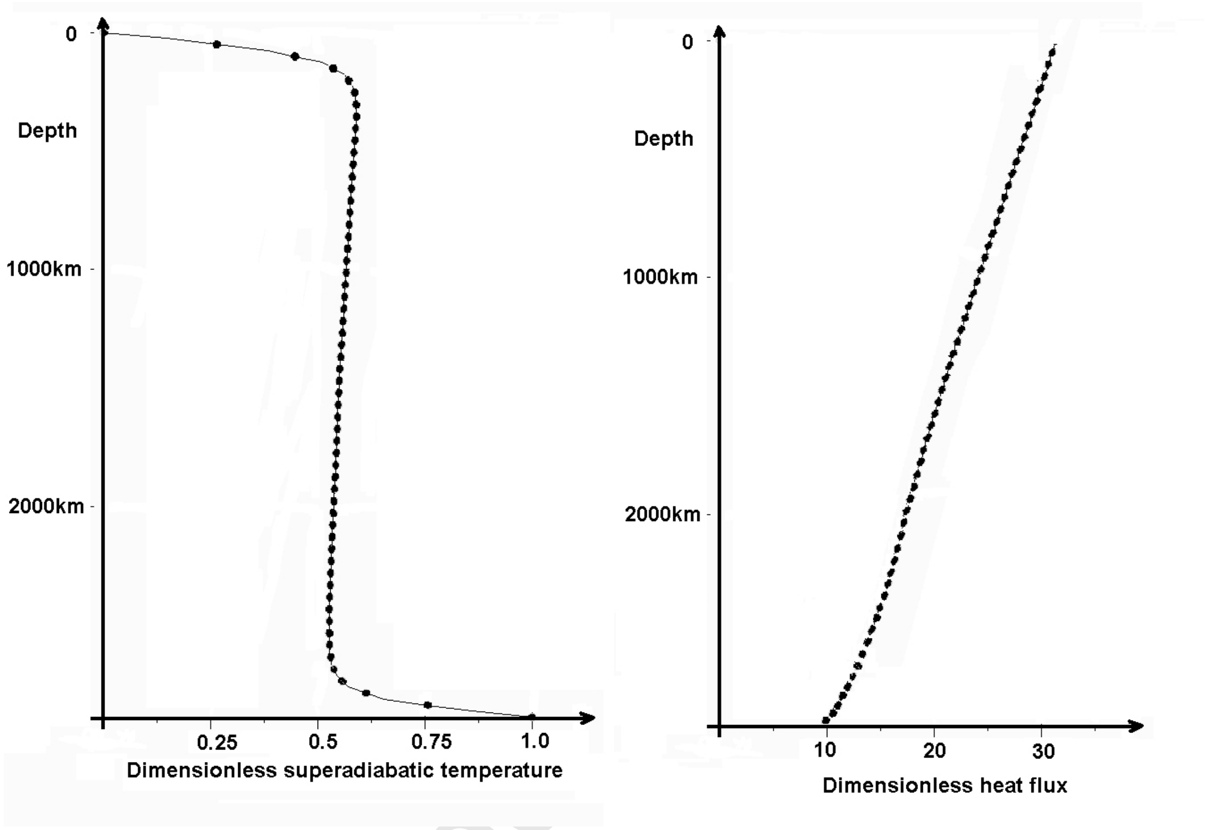

Fig. 1. The calculated dimensionless radial superadiabatic temperature distribution T(r) and heat flux q(r) for steady state free convection.

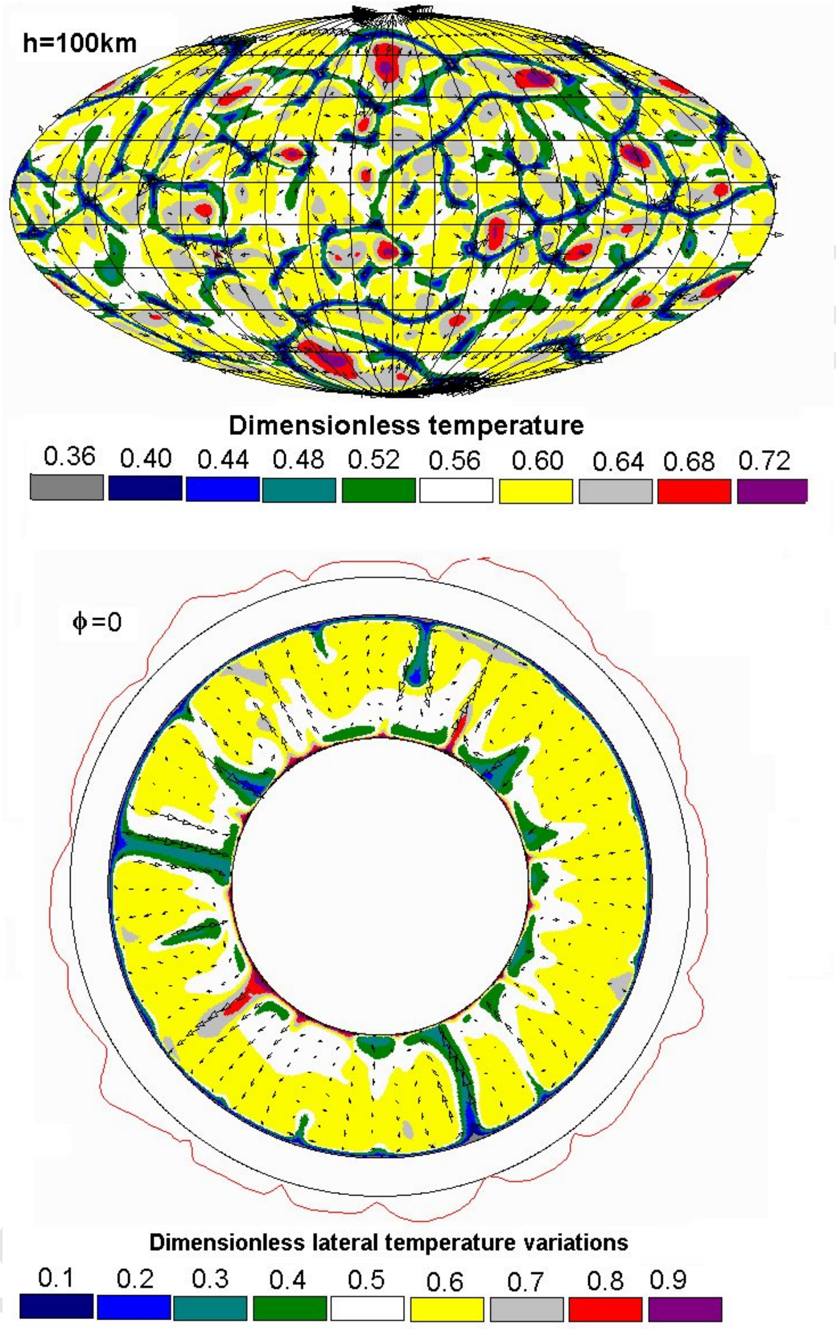

Fig. 2. Free steady state mantle convection for $Ra_R=10^7$ and additional internal heating H=100. The model grid is 58x144x216. Dimensionless lateral temperature variations are shown in color. Dimensionless mantle velocities are shown by arrows (maximal value is equal to $2 \cdot 10^3$). (Upper) - at a depth of 100km; (bottom) - for a sample cross-section, φ=0. The curve around the cross-section demonstrates the calculated relative surface heat flux q(θ).

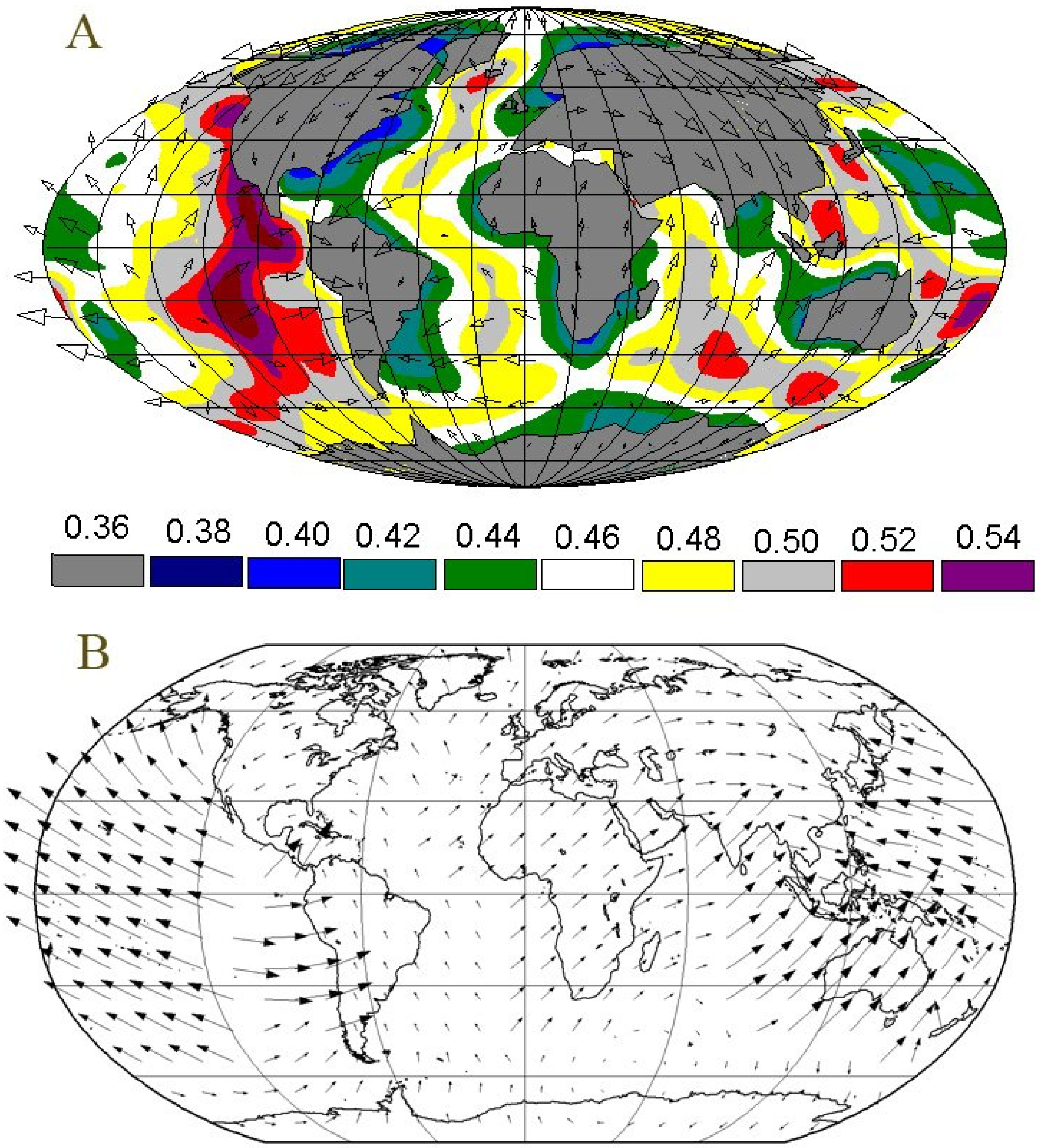

Fig. 3. (A) dimensionless temperature at a depth of 100km and mantle/continental velocities calculated after the initial time step for the mantle with the initial lateral temperature variations obtained from tomography data. Temperature is shown by colour, velocities – by arrows with the maximal value (dimensionless) - $2\cdot 10^3$. (B) Plate velocities according to the REVEL model (Sella et al., 2002) based on GPS data. Maximal value is equal to 10 cm/y.

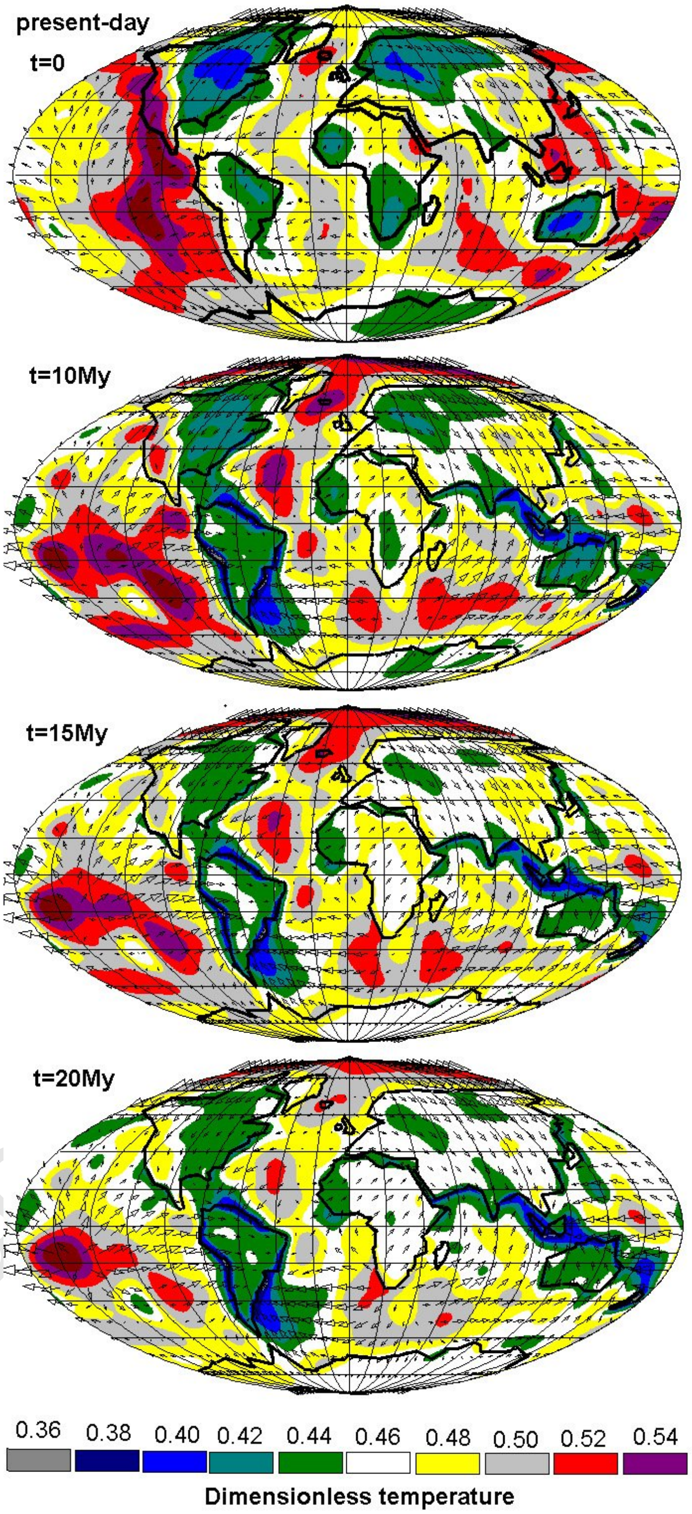

Fig.4. Evolution of the mantle convection pattern. Calculated temperature at a depth of 100km and mantle/continental velocities at the first stage of evolution and at time t=0, 10, 15 and 20 My. Temperature is shown by colour, velocities – by arrows with the maximal dimensionless value $2 \cdot 10^3$.

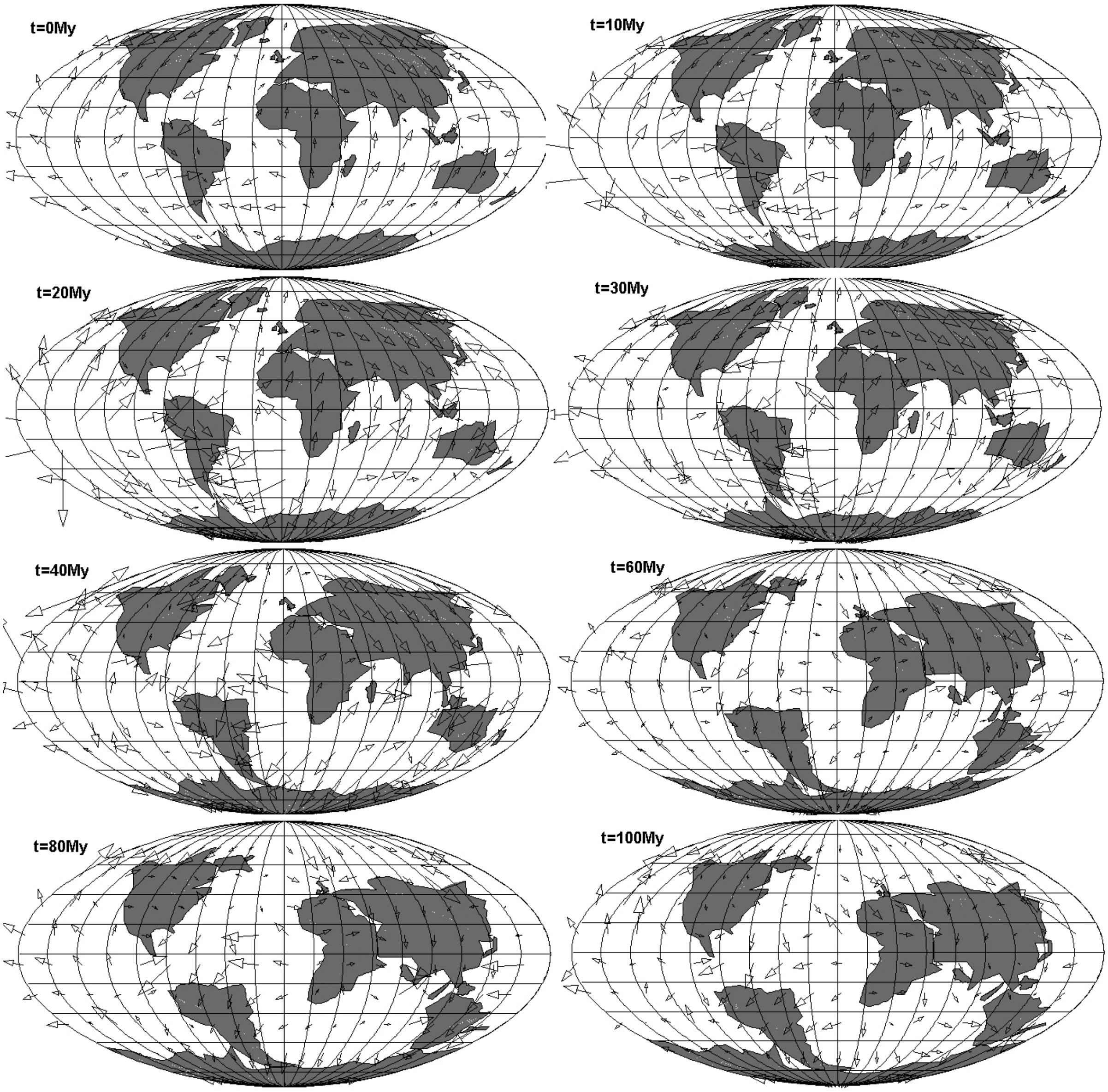

Fig. 5. Positions of the continents and mantle velocities for different time steps from 0 through 100 My. Dimensionless velocities are shown by arrows (maximal value $v_{max}=2\cdot10^3$).

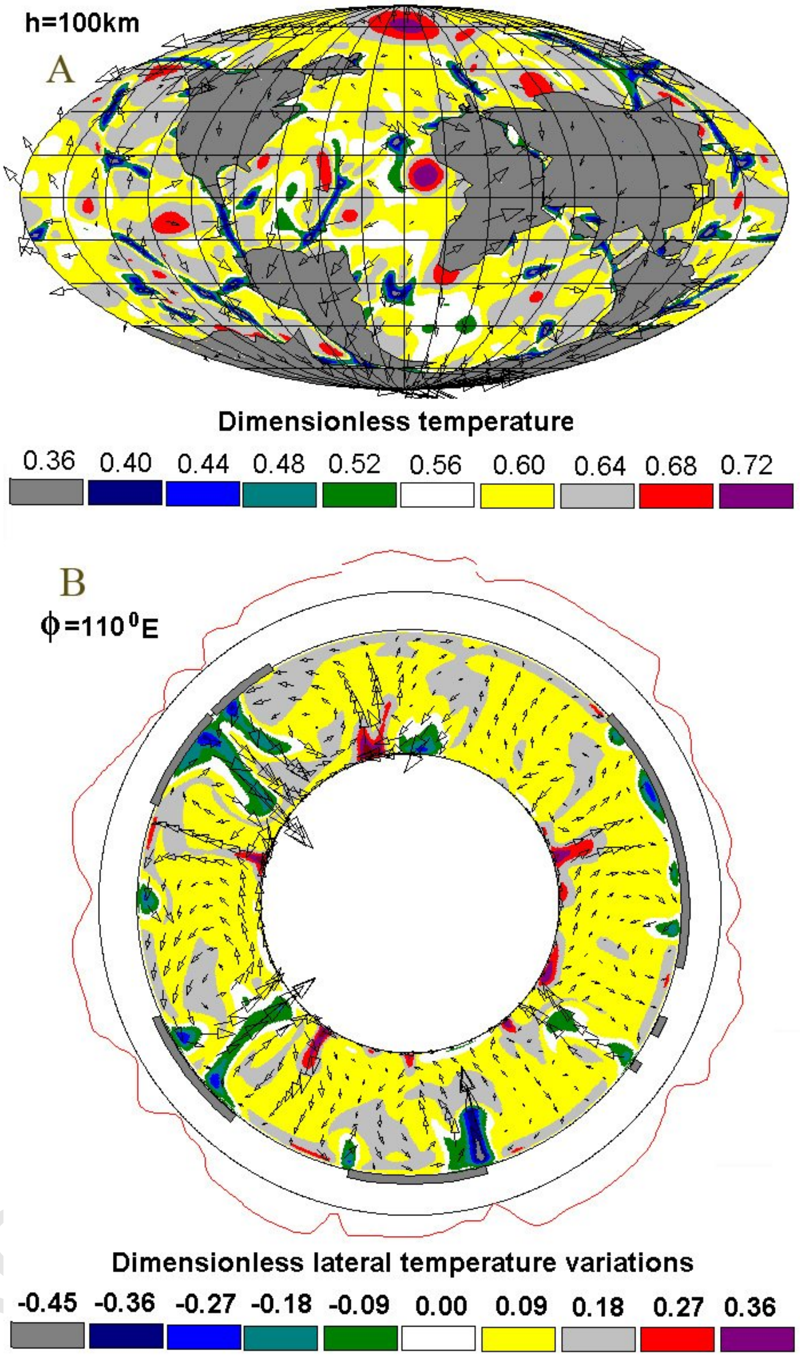

Fig. 6. (A) Mantle convection pattern and positions of the continents at a time of 100My. Dimensionless temperature at a depth of 100km is shown by colour, dimensionless velocities – by arrows (maximal value $v_{max}=2\cdot10^3$). (B) The same for a sample cross-section. The red line around the sphere represents the calculated heat flux (grey – reference level).

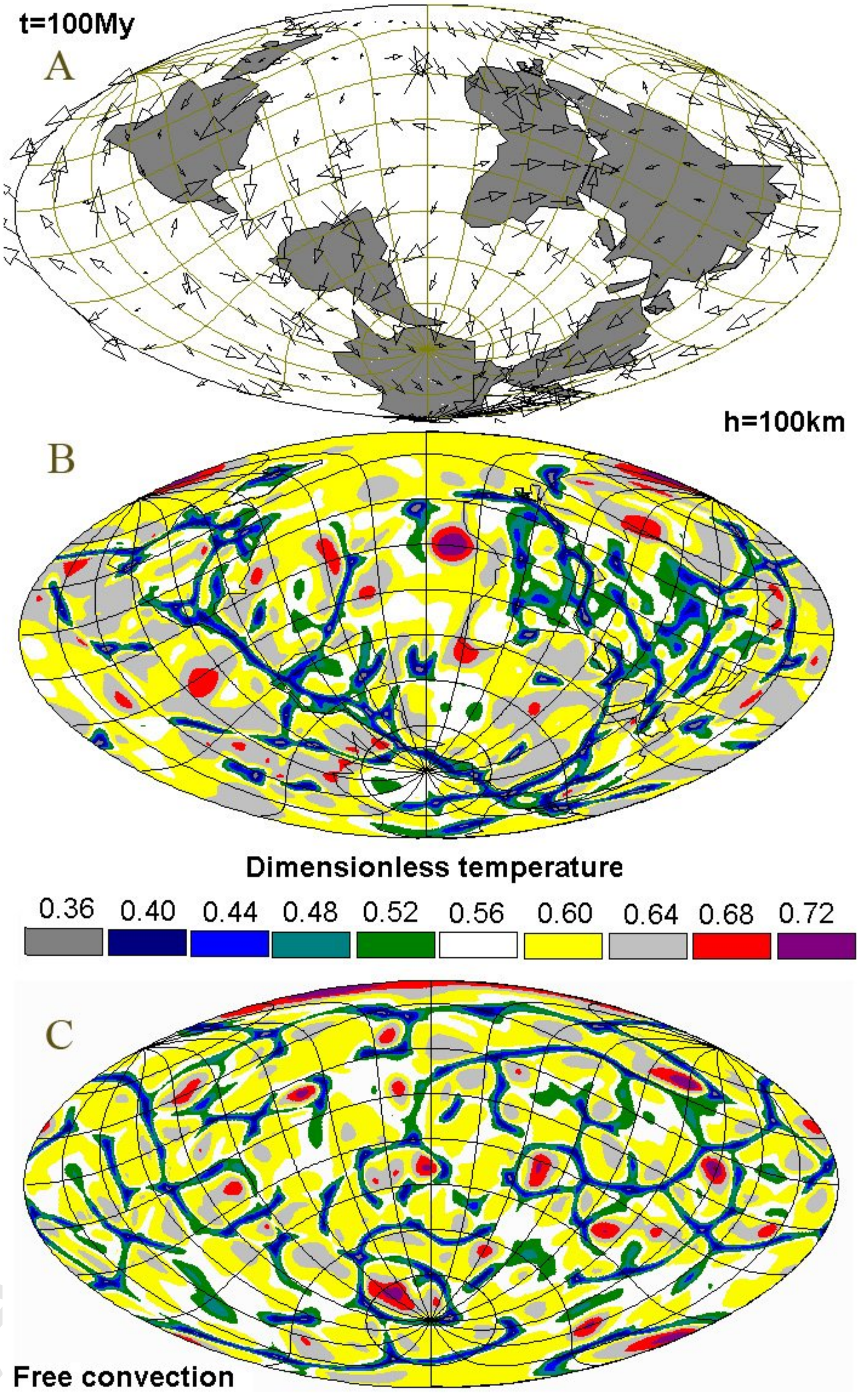

Fig. 7. (A and B) The same pattern of the mantle convection as in Fig. 6 shown from another point of view (from the southern hemisphere). (C)-free mantle convection from the same initial condition, but without continents.