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Short Communication

Numerical Experiments on Convection in a Chemically Layered Mantle

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Key words: Mantle convection – Finite element model – Chemical heterogeneity – Boundary layer – Mantle temperatures.

Introduction

It is still an unresolved question whether the convection cells connected with lithospheric plate motion are restricted to the upper mantle or involve the whole mantle. In the last few years a number of papers have appeared which favor whole mantle convection because: the viscosity of the lower mantle seems to be only low or moderate (Davies 1977); it can explain large-scale heterogeneities in the lower mantle (O'Connell 1977); it can explain the D″-layer as a thermal boundary layer of whole mantle convection (Elsasser et al. 1979); it ensures reasonably high temperatures in the upper mantle (Kopitzke 1979).

The opponents of whole mantle convection argue that the temperature would be too low at the core-mantle boundary (Jeanloz and Richter 1979) or they deal with the stress distribution and seismic energy release in descending plates above the seismic cutoff at 720 km depth (Richter 1979). Formerly it was often assumed that high rigidity of the lower mantle or the hindering action of phase transitions would inhibit whole mantle convection. Now the idea of a chemically heterogenous mantle is preferred. Some authors consider the 650-km discontinuity to be a pure chemical change (Liu 1979), or possibly a phase boundary with a superimposed increase of iron content (Anderson and Jordan 1970), while others exclude any chemical differences (Watt and O'Connell 1978). Geochemical arguments may also support different chemical reservoirs in the mantle (O'Nions et al. 1979). Sammis (1976) found by marginal stability analysis that convection in separated layers is the prefered mode, if the density difference is more than 0.1%. However, since mantle convection is vigorous and far from the marginal stability case, it is likely that the limit must be higher. Because the viscosity of the lower mantle can hardly be assumed to be extremely high, two separate layers of convection in the upper and lower mantle must develop if there is a substantial chemical boundary.

In order to investigate the dynamics and temperature distribution of two-layer mantle convection, such a boundary is introduced into a dynamical finite element model of convection.

Model Design

A detailed description of the model is given elsewhere (Kopitzke 1979, hereafter cited as paper 1; Christensen 1980, paper 2). Bicubic and biquadratic spline functions with Ritz-Rayleigh and upwind finite element techniques are used to evaluate the steady state flow and temperature field. Position-dependent viscosity, thermal diffusivity and expansivity, as well as inner heat production, bottom heat flux, frictional heating and adiabatic gradient are included. An advantage of this model is the realistic simulation of the upper thermal boundary layer which behaves mechanically and thermally like a nearly rigid lithospheric plate. The depth of the (plane) model is restricted to 1,750 km, in order to adjust the volume of the cell to the volume of a spherical sector of same surface area and 2,900 km depth extent. By slight variations of the Rayleigh-number (which is in the order of 10^7), the overturn time of the model lithosphere is made to equal the mean overturn time in the earth's plate system. This ensures realistic simulation of the heat balance in the mantle.

The chosen viscosity distribution exhibits a maximum of $>10^{25}$P on top of the lithosphere, a minimum of 10^{21}P at 200 km, rises again with increasing depth, reaching 10^{23}P at 500 km, and remains fairly constant (about $4 \cdot 10^{23}$P) in the lower mantle. Weak zones at the active margins of the lithosphere ensure its plate-like behavior (paper 1). The chemical boundary at $\simeq 650$ km depth is represented by a point-chain, as in a previous investigation, where a phase boundary was modelled (paper 2). However, its depth is not controlled by temperature as in the case of phase transition, but it is shifted up or down by material flow perpendicular to the boundary. There is a negative feed-back on the flow because of the buoyancy forces associated with boundary distortion, and usually steady state with zero flux through the dividing line is achieved (for a detailed description of buoyancy calculation in the finite element scheme see paper 2).

Results

In the first model the whole density increase at the 650-km discontinuity ($\Delta\rho/\rho = 10\%$) is assumed to be due to chemical change, according to the petrological model of Liu (1979) with pure perovskite in the lower mantle. The initial temperature distribution was chosen in order to establish two cells with opposite circulation in the upper and lower mantle. Steady state was attained easily by iteration; the results are demonstrated in Figs. 1 and 2. The upper cell looks very similar to a previous model with upper mantle convection only and a no-slip lower boundary (paper 1),

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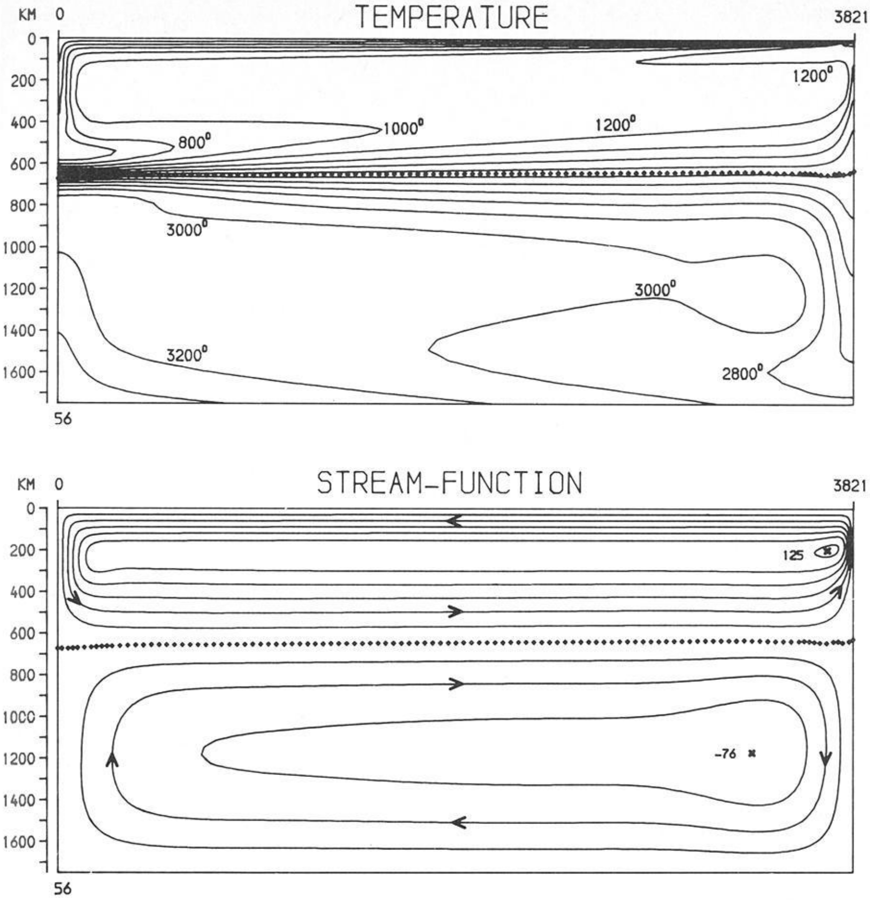

Fig. 1. Two layer convection, $\Delta\rho/\rho = 10\%$, temperature (contour interval 200°) and stream function (contour interval 20 dimensionless units). The chemical boundary is marked by dotted line. The spreading center is on the right, the subduction zone on the left

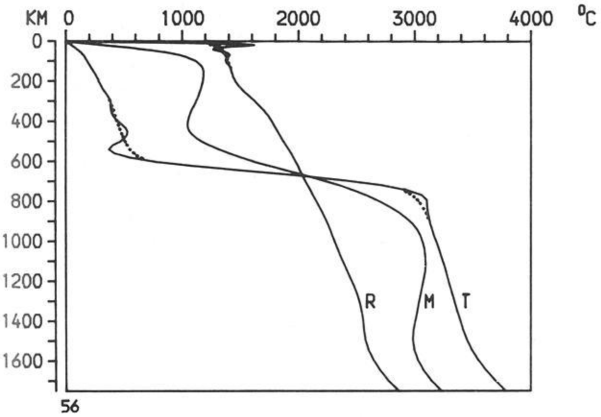

Fig. 2. Geotherms beneath the ridge (*R*), in the middle of the cells (*M*), and at the trench (*T*). Numerical over-swing effects are smoothed by *dotted lines*. The middle geotherm is representative for most of the length of the cell except narrow marginal regions

in terms of the flow as well as the temperature distribution. In the lower cell the flow is slower than in the upper one – 1 cm/a maximum velocity compared to 2.2 cm/a plate velocity and up to 10 cm/a in the ascending current beneath the spreading center. The temperature distribution in the lithosphere is in accordance with current estimates (about 1,100° C at its bottom), however, below the plate discrepancies occur. The mean temperature at the depth of the olivine-spinel transition (370 km) is 1,060° C – too far below the temperature estimates of 1400° C deduced from the properties of the phase transition (Graham 1970; Akimoto et al 1976). The chemical boundary is shifted down beneath the subduction zone and elevated beneath the spreading center by some tens of kilometers. At the dividing line a huge thermal boundary layer has developed – a result of the fact that 75% of the total surface heat flux must pass this boundary by thermal conduction. The mean temperature difference over the depth range from 500 to 800 km depth is 1,600°. Accordingly, temperatures are much higher in the lower mantle (Mean value 3,000° C, 3,200° C at the bottom7).

Another model, with a chemical density difference of only 3% (e.g., due to increased iron content of the lower mantle) failed to reach steady state. Wavy oscillations of the boundary established themselves in the region of diverging flow beneath the spreading center. Slight traces of these waves can also be seen in the first model (Fig. 1). It seems possible that smaller lumps (< 10 km) of the heavy material can be torn off and transported to the surface by the fast rising flow in the spreading axis, provided that the chemical density difference is only a few percent.

In a third experiment the chemical boundary was suddenly removed from the first model after steady state had been established. Immediately the descending plate penetrated the lower mantle and soon the separate lower cell was completely superceded by whole mantle convection. Thus the two-layer mode of convection seems extremely inferior without chemical heterogeneity.

Lastly, the experiment was carried out in reverse: a chemical boundary ($\Delta\rho/\rho = 10\%$) was introduced into a steady state whole mantle convection cell. Two separate layers of convection developed, the direction of plate motion was the same as before, however, the circulation in the lower cell was anticlockwise at the beginning, just as in the upper cell – in the opposite sense to the first model in Fig. 1. During this stage, the distortion of the chemical boundary was rather high – its depth varied by up to 180 km over the length of the cell. Yet the lower cell was not stable, a third 'embryo cell' with clockwise rotation appeared in the lower mantle and gradually superseded the other one. Finally everthing became the same as in Fig. 1.

Conclusions

1. Provided the viscosity in the lower mantle is not much higher than 10^{24} P (which is unlikely), a chemical boundary seems necessary for two-layer convection.

2. Two-layer convection can easily produce temperatures at its lowermost boundary, which are in the range of current estimates of the temperature at the core-mantle boundary (Stacey 1977), whereas the whole mantle model displays considerably lower ones. However, as discussed in paper 1, the latter is very probably due to several model idealizations.

3. It seems difficult for two-layer convection (as well as for upper mantle convection alone) to yield reasonable high temperatures in the upper mantle, while the whole mantle model was successful (paper 1). However, it is possible that special model assumptions or idealizations are responsible.

4. The very vigorous thermal boundary layer between the two cells would have a striking influence on a number of physical properties: seismic velocities (only moderate influence) and Q-factor, electrical conductivity, viscosity. For example: if Sammis et al. (1977) are right, and the activity energy for creep does not increase by more than 10 kcal/mol at the seismic discontinuities in the mantle, the viscosity would decrease by 8 orders of magnitude from the top to the bottom of the boundary layer. Temperatures as high as 2,800° C at 800 km depth would probably cause extensive partial or total melting. However, the temperature difference through the layer may be smaller than in the model, due to a smaller amount of heat sources in the lower mantle and core or due to higher velocity in the boundary layer. Furthermore the temperature effect on these properties may by obscured by the superimposed chemical effect. Nevertheless, if any boundary layer exists, it is not easy to understand why it has not yet been detected by its influence on seismic velocity or by the drop in Q, η, or σ.

5. If the coincidence between this model and the real mantle is not too weak (due to model idealizations or incorrect assumptions about model parameters such as viscosity or heat sources), a slight preference for whole mantle convection can be deduced.

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Book Reviews

H. Stiller und M.-P. Volarovich, editors. Physical Properties of Rocks and Minerals Under Extreme *P*,*T*-Conditions. Akademie-Verlag: Berlin, 1979, pp. 233, 146 figures, DDR-M 38,–.

This book contains a selection of papers summarising results obtained by working group 1.11 of the commission of the Academies of Sciences of Socialist Countries on the complex problem "Planetary Geophysical Studies". As is emphasised in many of the papers, knowledge of the behaviour of rocks and minerals with increasing pressure and temperature is essential to our understanding of many geophysical and geological processes. However, the title of this book is very misleading: the thermodynamic conditions described are in general $P \lesssim 25$ kb (or more often $\lesssim 15$ kb) and $T \lesssim 500$ °C. In view of the current capabilities of quasi-hydrostatic compression apparatus with internal heating ($P \sim 150$ kb, $T \sim 1{,}200$ °C) and diamond anvil cells ($P \sim 1$ Mb, or $P \sim 260$ kb, $T \sim 2{,}000$ °C with laser heating) these values can hardly be considered extreme. The scope of this book is thus limited to properties relevant to crustal and uppermost mantle processes, and does not include discussions of mantle phase changes, the nature of the low velocity zone and its relation to magma generation, or the lithosphere-asthenosphere boundary. There are also no references to shock was studies, which are certainly relevant to extreme *P*,*T* behaviour, nor is there any discussion of anelastic behaviour such as attenuation, although some determinations of effective viscosity are included.

The book is divided into six chapters, as follows:

1. Elastic Properties

Six papers, including two describing the apparatus used, covering determinations of ultrasonic velocities and elastic moduli in various rocks and minerals at pressures up to 15 kb and temperatures to 400 °C. The sixth paper summarises the results of a number of studies and applies them to determining lithological models for the crust in the Ukralnian shield. There is no discussion of the problems associated with ultrasonic velocity determinations, especially in small samples, nor is there any reference to Brillouin scattering techniques.

2. Deformational Studies

This section includes a theoretical analysis of systems of cracks in a stress field, a study of residual strain in granites under differential stress, and measurements of physical properties including strength, velocity and viscosity of various rocks which are then compared with seismic data for the crust in Armenia.

3. Electrical Properties

The first paper contains much useful information on the resistivity of upper mantle rocks and minerals ($P \lesssim 20$ kb, $T \lesssim 650$ °C). The next two describe electrical spectroscopy of second order phase transitions, and the surface conductivity and polarisation of rocks at high pressures. Unfortunately there is no detailed discussion of the relationship between high conductivity layers and partial melting in the light of the experimental results. The final paper, on the electrical conductivity of MSS structures in the Fe-Ni-S system, is of relevance to the structure of the earth's core, but the range of pressure (< 57 kb) and temperature (< 300 °C) makes large extrapolations necessary, and there is no reference to other determinations of the Fe-Ni-S phase diagram to 100 kb or to the implications of recent shock wave measurements on iron and iron sulphides.

4. Magnetic and Thermal Properties

Three papers covering the magnetic properties of iron cherts, the Hopkinson effect in magnetite and titanomagnetite, and the application of pulse measuring techniques to determination of thermal conductivities.

5. Physical Properties of Lunar Rocks

A single paper covering resistivity, thermal expansion, Young's modulus and compressibility of several types of rock.

6. Geophysical Consequences and Applications

The first paper contains a formulation describing the effect of porosity and pressure on wave velocity and its bearing on earthquake prediction. The authors address the difficult problem of the interaction between crack systems, but make no comparison between their results and those of other theories, and omit a discussion of the effects of varying crack aspect ratios. The statement that liquid movements are unimportant at pressures over 100 MPa is open to dispute; if true then the dilatancy-diffusion theory used to explain many precursory effects would not be valid for depths $\gtrsim 3$ km. The remaining papers discuss the use of laboratory measurements in the interpretation of DSS profiles, the connection between high pressure and planetary physics (including accretion) and a mathematical model for the occurrence of sudden stepwise phase transformations as an explanation of deep earthquake activity.

The scope of the work described here is extremely broad: unfortunately the individual papers are often too detailed to provide a clear overview or too general, perhaps because of length restrictions, to satisfy the specialist. It will mainly be of interest to those already working in the field of high pressure geophysics, for whom it provides a useful review of Soviet and eastern European literature (references to recent work in America and Japan are sparse) and those involved in the interpretation of deep electrical or seismic sounding data. Unfortunately the appeal of the book is lessened by the large number of printing errors, the poor reproduction of photomicrographs and the quality of the English which is frequently bad enough to obscure the meaning completely.

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