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REMARKS ON MANTLE CONVECTION AND THE EARTH'S EVOLUTION

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ABSTRACT

A few unresolved problems of the convection in the Earth's mantle are considered. It is discussed from both the geophysical and geochemical point of view whether the flows reach from the core-mantle boundary up to the oceanic lithosphere or whether the bulk convection is taking place in two or more layers, apart from a few plumes. The seemingly solved question whether the Earth is essentially cooling down in the course of its history is raised anew. A simple model of the thermal development of the lower mantle is presented and verified.

ZUSAMMENFASSUNG

Es werden einige ungelöste Probleme der Konvektion im Erdmantel erörtert. Sowohl vom geophysikalischen als auch vom geochemischen Standpunkt aus wird erwogen, ob die Strömungen von der Kern-Mantel-Grenze bis zur ozeanischen Lithosphäre reichen oder ob die allgemeine Konvektion in zwei oder mehreren Schichten stattfindet, wenn man von einigen Plumes absieht. Die scheinbar gelöste Frage, ob sich die Erde im Laufe ihrer Geschichte im wesentlichen abkühlt, wird erneut aufgeworfen. Ein einfaches Modell der thermischen Entwicklung des unteren Mantels wird vorgestellt und verifiziert.

1. INTRODUCTION

Subsolidus convection in the Earth's mantle is nowadays thought to be the decisive driving force behind many endogenous geological phenomena. In particular, the long-range, slow and chiefly horizontal movements of the oceanic and continental lithospheric plates, but also orogenesis, the spatial and time distribution of magmatism and the magnitude of the heat flow through the ocean floors can be explained best by means of convection models. Convection is heat transport that is tied to mass transport. With the Earth's mantle, convection due to differences in density comes into consideration. If the differences in density are the result of differences in the temperature, we deal with thermal convection. It has most probably been predominant for the major part of the Earth's history up to the present time. The heating is caused by the radioactive decay of U, Th and K. If the density differences which cause the driving force of convection are due to the chemical composition, we speak of compositional convection.

In place of further introductory remarks, we refer to a few reviews on mantle convection (Busse, 1978; Walzer, 1981; Boss, 1983, Spohn, 1984; Czechowski, 1986) and one on the kinematics and driving mechanisms of the lithospheric plates (Jacoby, 1985). In the present considerations, the main emphasis will be on some partially unresolved problems of mantle convection. A question that may be put first is:

2. DOES WHOLE-MANTLE CONVECTION, LAYERED CONVECTION OR UPPER-MANTLE CONVECTION PREVAIL?

2.1. Geophysical Considerations

Here, two partial questions are whether the slab can penetrate the 670-km discontinuity (Dziewonski and Anderson, 1981) and whether this discontinuity is a phase transition or a change in the chemical composition. Toksöz et al. (1971) developed a model in which the downgoing slabs of the oceanic lithosphere, which are situated deeper than 700 km, are no longer brittle enough to accumulate the stress for earthquakes. Beneath various subduction zones such as under the Kuril-Kamchatka slab up to a depth of 1000 km (Craeger and Jordan, 1984), beneath the Caribbean Sea up to a depth of 1400 km (Jordan and Lynn, 1974) and beneath the Tonga-Fiji arc (Engdahl, 1975) anomalously high seismic velocities have been found. Schubert et al. (1975) have proved that phase transitions do not prevent the penetration of convection flows. Ringwood (1979) has affirmed that there is sufficient evidence for a large degree of chemical homogeneity throughout the mantle. Cathles (1975) claimed that the lower mantle has about the same low viscosity as the upper mantle. All of this has led to the assumption that the aforementioned zones of increased seismic velocity constitute the continuation of the subduction slabs into the lower mantle and to the computation of a multitude of models for whole-mantle convection.

Many facts, however, go counter to an interpretation of the 670 km discontinuity as a phase boundary: Seismic observations (Sobel, 1978; Husebye et al., 1977) imply that about 5% of the incident seismic energy (at 1 Hz) is reflected at this discontinuity. Lees et al. (1983) have computed the increase in density and seismic velocities with depth, assuming olivine, pyroxene, garnet and olivine-plus-garnet model mineral assemblages. For chemically homogenous mantle models, reflection coefficients smaller than

0.5% were obtained. The variability of the depth of the discontinuity also goes counter to a phase transition. Jeanloz and Thompson (1983) have found a discontinous reaction in the olivine component of the mantle only at a pressure corresponding to a depth of 400 km, but not at 670 km. The transition from γ-spinel to a silicate-perovskite assemblage does not satisfy the sharpness of the 670 km discontinuity. Jeanloz and Thompson (1983) concluded from this that either a not yet observed, univariant reaction or a chemical discontinuity must be present. Loper (1985) has proposed a hypothetical non-equilibrium garnet-to-perovskite phase transition to explain the sharpness of the discontinuity. It can be shown, however, that the flow velocities involved are too low to create such effects.

The way in which real mantle convection takes place also depends to a large degree on the magnitude of the viscosity in the lower mantle. Its order of magnitude must be in the range between 10^{21} and 10^{25} Pa•s. The correct value is disputed (for more details, see Walzer (1981), Section 4.1). From the isostatic uplift of differently sized areas (Laurentia , Fennoscandia, Lake Bonneville) after the post-glacial relief, Cathles (1975) and Peltier and Andrews (1976), using different mathematical methods, have arrived at very similar results. Cathles' (1975) model is as follows: Depending on the area in question, the thickness of the elastic lithosphere varies between 70 and 150 km. A 75 km thick low-viscosity layer with $4 \cdot 10^{19}$ Pa•s lies beneath it. The remaining upper mantle has $(1.0 \pm 0.1) \cdot 10^{21}$ Pa•s. For the lower mantle, Cathles has found $(0.9 \pm 0.2) \cdot 10^{21}$ Pa•s. The entire mantle was assumed to be a Newtonian fluid.

It is typical of the representatives of whole-mantle convection that they assume at the same time for the lower mantle relatively low viscosity values and a chemical composition similar to that of the upper mantle. These two assumptions remain incompatible with the observation that focal mechanisms at downgoing slabs, which reach only to a small to medium depth, indicate tension (down-dip extension mechanisms), whereas slabs reaching a depth of 600 to 700 km show compression mechanisms over their entire length (Isacks and Molnar, 1971). A classification of the 39 modern subduction zones into 7 semiquantitative strain classes is given by Jarrard (1986). The conclusion that suggests itself, namely that the movement of the slab at a depth of 700 km is braked by buoyancy and (or) by downward increasing viscosity is in contradiction to the first or the second mentioned premise for whole-mantle convection. The only whole-mantle convection model I know which avoids this difficulty is that put forward by Loper (1985). He assumes that the idea that mantle convection is driven by internal heating is wrong. He introduces the hypothesis that, except for the generally accepted cooling at the Earth's surface and the resulting thermal boundary layer, which is identical with the oceanic lithosphere, the mantle is essentially heated from the core at the core-mantle boundary. According to him, the layer D'' is a hot thermal boundary layer, from which the plumes eat, like in a hose, through the entire mantle to generate hot-spot volcanism on the surface of the Earth (Stacey and Loper, 1983; Loper and Stacey, 1983), with the lower mantle remaining essentially undisturbed and having a pristine chemical composition.

The aforementioned distribution of the focal mechanisms in the downgoing slabs is interpreted by most representatives of whole-mantle convection as being due to the fact that the downgoing slab is dissolving there, while its parts are sinking still further. However, in such a case it remains inexplicable why seismicity so sharply discontinues downwards because, according to this hypothesis, the slab is vanishing downwards so that the earthquakes should become progressively rarer and weaker in the downward direction. Instead, however, seismicity shows a distinct maximum between 500 and 700 km and suddenly disappears below a depth of 700 km (Billington, 1978; Hanuš and Vaněk, 1979). It appears that the change in the character of the focal mechanisms can only be explained by the fact that, up to medium depths, the downgoing slab is additionally sinking through its own weight, i.e., it is pulled, while it meets resistance at greater depths. One might assume that this resistance is due to an increase in viscosity (in particular at the 670-km discontinuity), but also to the buoyancy in a denser medium in the case of a chemical stratification of the mantle, most probably, however, to a combination of these two factors. Richter (1979) comes to the conclusion that the fact that an earthquake has never been observed in the lower mantle indicates that the downgoing slab does not penetrate into it because of chemical stratification. The approximately horizontal lower end of the downgoing slab located at the New Hebrides at a depth between 600 and 700 km also points in this direction. That is, the corresponding distribution of the seismicity shows that the slab obviously has not been able to penetrate to a greater depth (Isacks and Molnar, 1971; Giardini and Woodhouse, 1984). On the other hand, the problem of the viscosity of the lower mantle is not resolved. Using the secular variation of the gravitational harmonic coefficient J_2, Yuen and Sabadini (1985) found two families of solutions. The first solution is a nearly constant viscosity, namely 10^{21} Pa•s. The second solution curve is, however, rising from 10^{23} Pa•s at a depth of 670 km to 10^{25} Pa•s and more at greater depths of the lower mantle.

2.2. Geochemical Considerations

While the oceanic lithosphere has to be regarded as a thermal boundary layer, the continental lithosphere can be designated as a chemical boundary layer (Jordan, 1981). The latter one is obviously thicker than has been previously assumed. 3200 to 3400 Ma old garnets in diamonds prove that, at least in South Africa, the thickness of at least 175 km of the continental lithosphere has not changed since that time (Richardson et al., 1985; Richter, 1985). In the course of time, the continents grow through orogenesis, and from the recent period we can essentially distinguish between the subduction type and the continental collision type. Formerly, it was somewhat disputed to which degree the granitic magmas created during orogenesis are the result of recycling or are newly produced from the mantle. Two extreme standpoints put forward were as follows: a) the continent is continuously formed from the mantle, i.e., the volume of the continents is growing linearly (Hurley et al., 1962; Moorbath, 1977); b) a primitive continental crust differentiated during the Archean and has then been repeatedly reworked in a circle consisting of erosion, sedimentation, metamorphism and anatexis (Armstrong, 1981). The truth appears to be in between. Allègre and Jaupart (1985) concluded from the budgets of isotopic systems that the mean age of continents is between 2.1 and 2.4 Ga. Turcotte and Kellogg (1986) arrived, through investigations of the mass balance based on samarium-neodymium and rubidium-strontium data, at a mean continental age of (2.1 ± 0.7) Ga. The development of the Rb/Sr, Sm/Nd and of the $^{143}Nd/^{144}Nd$ ratios is proof of a significant share of reworking in young granites. On the other hand, many Precambrian granitoids stem from the mantle. Allègre and Jaupart (1985) concluded from this that the total mass of the continents has been developing since 4 Ga along a more or less sigmoidal curve, with a steady-state value having been reached now. The zero-point of the curve is given by the extraction age of oldest known rocks of maximally 4.1 Ga and the highest age of the zircons of 3.8 Ga. The fact that the share of the continental material in the sediments during the Precambrian was much smaller than in the Phanerozoic also speaks in favour of this continental growth curve. Windley (1977) and Sawkins (1984) have gathered a lot of data on lithophile element enrichment in the crust. In the Precambrian parts of the continents, the concentration of K, Th, U and Rb within one facies is decreasing with depth. For a constant depth level, the same is true of the transition from granulite to amphibolite facies in Precambrian gneisses.

A first significant conclusion for future models of convection with heating from within is that, due to the migration of the incompatible elements (in particular, U, Th, K) into the evolving continents, their concentration in the depleted mantle reservoir must have been approximately decreasing according to a sigmoidal curve (Allègre and Jaupart, 1985). This does not apply to another layer of pristine chemical composition in the mantle, whose existence will be advocated below. There is reason to believe that the thickness of the Precambrian portions of continents can be assumed to be constant in time (Richter, 1985). Therefore, the idea suggests itself to explain the decrease of the present average heat flow from 41 mW/m^2 for Archean portions of the continents to 51 mW/m^2 for Proterozoic portions of the continents (Chapman, 1985) in terms of an increase in the rate of migration of U, Th and K into the continental crust. The Archean crust is on the average depleted in U, Th and K relative to younger crust. The explanation suggested by Morgan (1985) is that the crust depleted in radiogenic heat sources is less susceptible to reactivation and that, therefore, the remaining Archean crust is a biased sample. In answer to this, it has to be said that orogenesis and its magmatism is tied to collision zones and subduction zones and obviously does not originally develop from the crust.

Secondly, in the future, one should combine the gravitational differentiation at the subduction zones and other aspects of compositional convection together with the thermal convection of the mantle (which includes the movement of its thermal boundary layers) in one model. The papers by Richter (1986), Ribe (1985), Brandeis and Jaupart

(1986), Longhi and Ashwal (1985) and Turcotte and Pflugrath (1985) can be considered as preliminary studies in this respect.

Now, basalts are to be the subject-matter of our considerations. They can be roughly subdivided into three classes, which differ from one another in their contents of incompatible trace elements. The term 'enriched' is to denote a high content of U, Th, lanthanides, Rb, Sr, K and P, and the term 'depleted' a low content. The basalts of the mid-ocean ridges (MORB) with a bathymetric depth greater than 2 km are depleted. Their composition is very uniform. This does apply both to the main components and to the trace elements. The basalts of the oceanic islands (OIB) and the continental flood basalts (CFB) are enriched, with the composition of the CFB's strongly varying. These observations, which have been presented here in a simplified way, have led to the concept of a chemical stratification of the mantle, with many authors considering the upper mantle to be depleted. O'Nions et al. (1979) concluded from the $^{87}Sr/^{86}Sr$ ratio that the lithosphere has been created through the differentiation of about one half of the mantle. For mid-ocean ridge basalts, a mean $^{87}Sr/^{86}Sr$ ratio of 0.7028 to 0.7030 has been found. If the entire mantle participated in the mass transfer to the continental lithosphere, we would obtain 0.7047. The correct value is obtained if only about one half of the mantle had been involved in the gravitional segregation. Jacobsen and Wasserburg (1979) constructed simple transport models for the formation of the crust from the mantle. Mass balance calculations showed that the continents have been formed through the differentiation of about 30 percent of the mantle. Thus, these studies also indicate that there must be different layers of convection in the mantle, namely, one for the lower and at least one for the upper mantle.

Next, using Figure 1, we shall consider the model suggested by Turcotte and Kellogg (1986). They assume that the isotopically and compositionally uniform MORB stems from the upper mantle (UM). Consequently, like most geochemists, they equate the depleted mantle reservoir, which is only approximately known in terms of its volume, with the UM. However, the UM is contaminated in some places through descending slabs, which are partially compositionally segregating when passing through the UM. They refer to other studies to explain the great homogeneity of MORB. The authors of these studies try to prove that the UM convection is a very effective stirring mechanism, which I am doubtful about. In their model, the mid-ocean ridge system randomly migrates over the UM. My view, however, is that the position of these ridges is generally caused by ascending convection flows which carry with them the material for MORB. According to Turcotte and Kellogg (1986), a very depleted lower oceanic lithosphere and a slightly enriched oceanic crust are formed at the mid-ocean ridges. As it is moving parallel to the Earth's surface, the oceanic crust is cooling down more and more and is pulled back through its own weight into the UM (slab pull). During the sinking of this slab, a U-, Th- and K-enriched continental upper crust and a continental lower crust, possibly Pb-enriched, are formed mainly through island arc volcanism. What is left of the two layers of the oceanic lithosphere forms a boundary layer at a depth of 670 km,

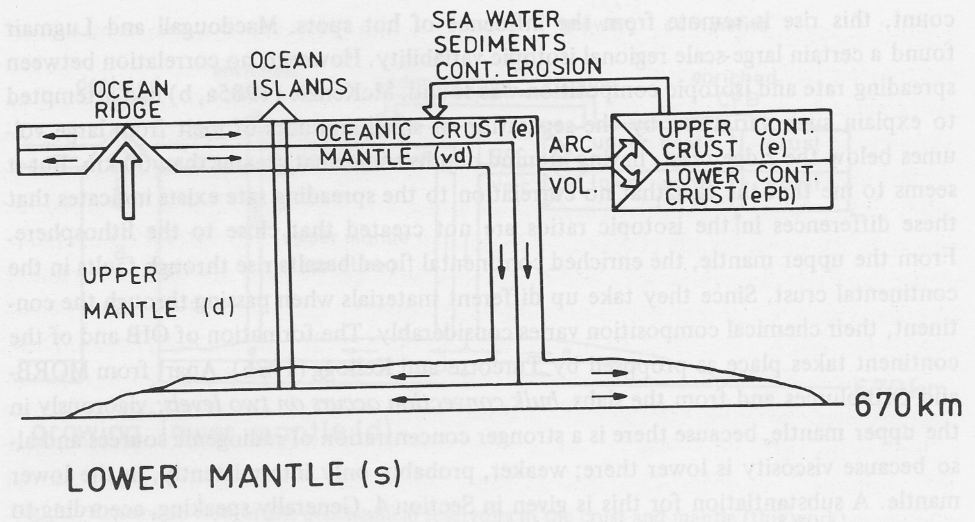

Fig. 1: Schematic view of the geochemical reservoirs in the crust and mantle according to Turcotte and Kellogg (1986).

from the enriched parts of which the OIB is rising in the form of plumes. I can subscribe to this part of the concept. This model for the generation of the OIB is related to the idea advanced by Hofmann and White (1982), apart from the fact that they assumed that the boundary layer is located at the core-mantle boundary. A substantiation why the Hofmann-White layer has to be localized at a depth of 670 km will be given after the next paragraph. Allègre et al. (1983) have also expressed doubts in the above assumption that the UM is the depleted mantle reservoir. The isotopic data of Hawaii and Iceland can be interpreted as the result of mixing between a depleted MORB source reservoir and an undepleted reservoir (Sun and Hanson, 1975; Poreda et al., 1986). Turcotte and Kellogg (1986) conclude from the Sm-Nd system that the ratio of the mass M_c of the continental crust to the mass of the depleted mantle reservoir is 0.010. On the other hand, the ratio of M_c to the mass of the upper mantle is 0.018. This means, however, that the mass of the depleted reservoir must be almost double that of the upper mantle. If we form the ratio of M_c to the mass of the lower mantle, we obtain 0.007, and this is much closer to the observed value of 0.010. *Therefore, I introduce the hypothesis that the lower mantle is the depleted mantle reservoir* and that, consequently, the upper mantle consists essentially of pristine mantle material (see Fig. 2). If we exclude D" from the lower mantle, we come even closer to 0.010. The very depleted lower oceanic lithosphere superimposes itself onto the 670 km discontinuity. In this way, it leads to a *growth of the depleted lower mantle.* Parts of it, possibly reheated by the mechanism proposed by Loper and Stacey (1983), rise under the mid-ocean ridges where they form MORB. Macdougall and Lugmair (1986) studied $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios on fresh glassy East Pacific Rise basalts. If we leave Easter Island and Galapagos out of account, this rise is remote from the influence of hot spots. Macdougall and Lugmair found a certain large-scale regional isotopic variability. However, no correlation between spreading rate and isotopic composition was found. McKenzie (1985a, b) had attempted to explain such variations by the separation of small amounts of melt from large volumes below the ridge crest, having in mind depths somewhat greater than 60 km. But it seems to me that the fact that no correlation to the spreading rate exists indicates that these differences in the isotopic ratios are not created that close to the lithosphere. From the upper mantle, the enriched continental flood basalts rise through faults in the continental crust. Since they take up different materials when passing through the continent, their chemical composition varies considerably. The formation of OIB and of the continent takes place as proposed by Turcotte and Kellogg (1986). Apart from MORB- and OIB-plumes and from the slabs, *bulk convection occurs on two levels:* vigorously in the upper mantle, because there is a stronger concentration of radiogenic sources and also because viscosity is lower there; weaker, probably only intermittently, in the lower mantle. A substantiation for this is given in Section 4. Generally speaking, according to the model which has been only roughly outlined so far, the evolution of the mantle is characterized by the fact that the continental lithosphere with a high heat source density and, below, a lower mantle with only a low heat source density are growing at the expense of an ever decreasing upper mantle with a pristine radioactive heat source density. Still, at collision zones, it is quite possible that a small fraction of the continental volume is dragged back into the mantle, which has been documented in Tibet. The amount of sediments supplied from the continental crust to the oceanic crust may vary considerably. Moreover, tectonic processes along modern convergent margins may by no means only lead to accretion but also to an erosion of the continental plate (von Huene, 1986). In some cases, both erosion and accretion have been observed on one and the same subduction zone. Thus, the new model (Fig. 2), though refined in some aspects and partly substantiated in a different way - essentially leads back to Stacey's old model (1969), at least with respect to the heat source distribution.

Hofmann and White (1982) have suggested that the subducted oceanic lithosphere sinks down to the core-mantle boundary where it forms a layer. Through heating from below, plumes are generated from this layer, which penetrate through the entire mantle and form OIB above. Although this is possible chemically, the author has his reservations in regard to the given depth of the layer since OIB hotspots appear to have been almost stationary for 100 to 200 Ma. An investigation of the hotspots of America, Europe, Africa and the Atlantic Ocean has yielded movements of about 3 mm/a only for some of them. The mutual position of the hotspots has changed only within a range of a few hundred kilometres (Morgan, 1983). Consequently, the place where the OIB plumes are generated must have a relatively fixed basis. However, it follows from the geomagnetic secular variation that the viscosity of the outer core must be very low, in any case significantly lower than 10^8 Pa·s; therefore, the outer core is out of the question for this. The Hofmann-White layer could be superimposed on the lower mantle, which, ac-

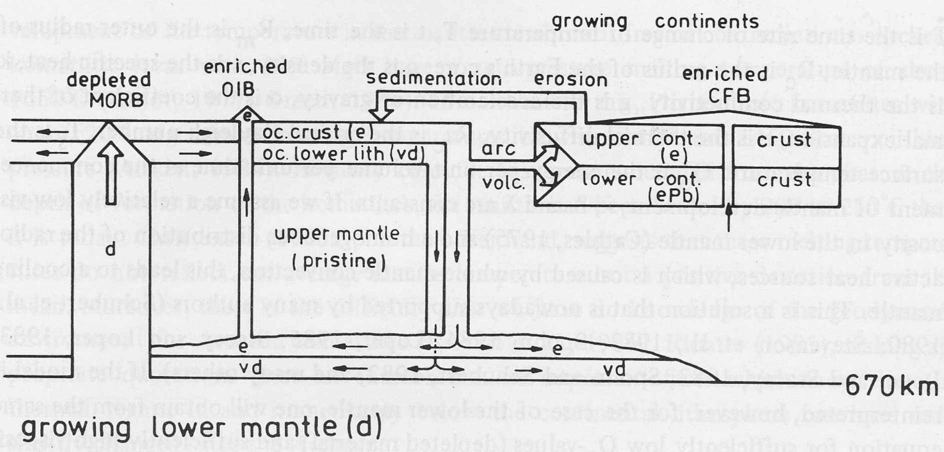

Fig. 2: Schematic view of the geochemical reservoirs in the crust and mantle (this work).

cording to Walzer (1974, 1978, 1981) can only episodically be in the state of bulk convection.

The assumption that the heavier product of differentiation lies *below* the pristine mantle suggests itself also because otherwise a global gravitational instability would be created. Another substantiation for two-layer bulk convection is given by Richter (1985): he calculates the observed heat flow if he assumes bulk earth uranium concentrations of 15 to 30 ppb in various two-layer models. This corresponds approximately to the published geochemical estimates. For whole-mantle convection, more than 35 ppb are necessary. Zindler and Hart (1986) concluded that if the lower mantle is chemically distinct and convectively isolated from the upper mantle, it does not have a primitive or undifferentiated composition.

3. DOES THE INTERNAL ENERGY OF THE EARTH MAINLY DECREASE OR INCREASE IN THE COURSE OF TERRESTRIAL EVOLUTION?

Schubert et al. (1980) have shown that the well-known exponential decay with time of the radioactive heat sources and the simultaneous assumption of a strict equality of heat generation and heat output to outer space, which is applicable in any geological era, leads to a contradiction. Using $\beta = 1/3$, the following equation is formed from equations (1) to (5) of Schubert et al. (1980).

$$\rho c \dot{T} + \frac{3R_m^2}{R_m^3 - R_c^3}\left(\frac{k^3\, g\, \alpha}{\kappa\, \bar{\nu}\, Ra_c}\right)^{1/3} (T - T_s)^{4/3}\, e^{-A/3T} = Q_o e^{-\lambda t}$$

$\dot{T}$ is the time rate of change of temperature T, t is the time, R_m is the outer radius of the mantle, R_c is the radius of the Earth's core, ρ is the density, c is the specific heat, k is the thermal conductivity, g is the acceleration of gravity, α is the coefficient of thermal expansion, κ is the thermal diffusivity, Ra_c is the critical Rayleigh number, T_s is the surface temperature, Q_o is the energy per unit volume per unit time at the commencement of mantle development, $\bar{\nu}$, A and λ are constants. If we assume a relatively low viscosity in the lower mantle (Cathles, 1975) and a homogeneous distribution of the radioactive heat sources, which is caused by whole-mantle convection, this leads to a cooling mantle. This is a solution that is nowadays supported by many authors (Schubert et al., 1980; Stevenson et al., 1983; Spohn, 1984; Loper, 1985; Stacey and Loper, 1983; Loper and Stacey, 1983; Spohn and Schubert, 1982, and many others). If the model is reinterpreted, however, for the case of the lower mantle, one will obtain from the same equation for sufficiently low Q_o-values (depleted material) and sufficiently high viscosity constants $\bar{\nu}$ solutions, smoothed in time, with initially strongly rising and, in the Phanerozoic, almost constant temperature (see Section 4).

The cooling-mantle models of the various authors frequently differ strongly from one another. For the ratio of surface heat flow to contemporaneous radiogenic heat production, Christensen (1985a) arrives at values in excess of 2, Davies (1980) obtains 2, Spohn and Schubert (1982) 1.4, McKenzie and Weiss (1975) 1.2, Turcotte (1980) 1.15 etc. Whereas Schubert et al. (1980) concluded that the initial mantle temperature has little effect on the final results, Christensen (1985a, b) has shown that the present thermal state is still significantly influenced by the initial mantle temperature. Both authors use whole-mantle models.

The theory of equilibrium condensation in the presolar cloud has led to the hypothesis that the terrestrial planets originally had a single uniform raw material. It was assumed that iron, silicates, Al_2O_3, CaO etc., originally were present in a homogenized mixture and that, at an early stage of the planet, the iron core and the silicate mantle differentiated from it. According to the theory, the potential energy released in the process strongly heats the Earth. The hot start models with a cooling earth are influenced by these models. Assuming a homogeneous accretion, however, one gets into difficulties, because water, carbon dioxide and other volatiles are too abundant on the Earth's surface. Also the early existence of life and of the hydrosphere points to heterogeneous accretion (Jagoutz et al., 1979; Wänke, 1981; Wänke and Dreibus, 1982). According to Wänke (1981), accretion commenced with highly reduced material free of volatile elements, but the other elements had C1 abundance ratios. During this early stage, the Earth's core also was formed, in which case its light component cannot be troilite but possibly metallized silicon. After about 2/3 of the Earth had formed, more and more oxidized material with moderate volatiles accreted and, finally, material with volatile elements also came to the primordial Earth. Mars obviously underwent a similar evolution (Dreibus and Wänke, 1985). These geochemical findings have encouraged the author to attempt a cold start model for the mantle. The observational results of the development of the temperature in the mantle are also not free of contradictions. Archean komatiitic lavas with 32% MgO have a zero-pressure melting temperature of about 1650°C (Arndt, 1977; Green, 1981). In contrast to this, Phanerozoic komatiites with about 20% MgO have melting temperatures between 1400 and 1450°C at vanishing pressure (Aitken, 1984). In many cases, it has been concluded from this that, at the depth of formation of the komatiites, the mantle had been hotter by 200 to 250°C than it is today. Christensen (1985a), on the other hand, suggests that these melting temperatures do not reflect the average mantle temperature but a positive deviation from the mean. Moreover, there is the difficulty that we have no knowledge of the geodynamic environment for Archean komatiites (Brévart et al., 1986). Richter (1985), making appropriate quotations gives an enumeration of important facts which show that the thermal regime in and immediately below the continental lithosphere can have only slightly changed since the Archean:

a) There is little difference in the conditions of metamorphism.

b) The normal crustal thickness of the Superior Province shows that the temperature gradient cannot have been much steeper in the Archean.

c) It follows from the examination of Archean granulites that the maximum height of the mountains at that time was close to that of Tibet today. Consequently, the increase in temperature with depth must have been very similar to that encountered today.

d) The investigation of 1200 to 1400 Ma old xenoliths in kimberlites shows that, at least under South Africa, the geothermal gradient cannot have changed significantly. The Archean geotherm is almost in complete agreement with the Cretaceous geotherm.

e) From Sm-Nd and Rb-Sr model ages of 3200-3400 Ma for garnets entrapped in the diamonds of 90 Ma old kimberlites, a lithosphere thickness of more than 175 km and a geothermal gradient almost unchanged since the Archean follows for South Africa.

The author, going further than Richter (1985), concludes from this that the mantle as a whole also has not significantly cooled down since that time. The aforementioned facts at least show that a cold start model involving strong heating at the beginning of terrestrial evolution and a rather uniform development of the temperature of the mantle during the main part of the Earth's history is not in contradiction to observations.

4. A SIMPLE MODEL OF MANTLE CONVECTION

In the model of convection in the lower mantle discussed here, by far not all aspects of the first three sections have been used. In particular the migration of radioactive elements has not yet been taken into account. Also convection in the upper mantle has been introduced only in a very general way. Important for the model is the assumption that the lower mantle is depleted. We assume according to Stacey (1969) that the lower mantle nowadays has a low specific heat production of 0.27 erg/g•a. In this case, the specific heat production of the upper mantle must be greater to yield the heat flow measured on the Earth's surface. Incidentally, it also follows from this that the upper mantle is continuously convectively flowing. Now, to become somewhat more independent of the special numerical assumptions, we shall use three competing models for the lower mantle: In Model 1, we assume, somewhat artificially, that the heat production density Q^* is constant in time; in the other two models, it exponentially decays according to the law of radioactive decay:

$$Q^* = Q_o^* \exp(-t^*/t_o^*), \tag{1}$$

where t^* is the time. The constants change according to the assumed mixture of the radioactive elements. For a chondritic composition, we have, according to McKenzie and Weiss (1975) $t_o^* = 2219$ Ma, for a composition according to Wasserburg et al. (1964) $t_o^* = 3248$ Ma. This is to apply also to our Models 2 and 3, respectively. We determine Q_o^* for Models 2 and 3 such that 4.5 Ga after the accretion of the Earth the specific heat output has fallen to 0.27 erg/g•a. Thus, we have

for Model 1 $Q_o^* = 4.2864 \cdot 10^{-8}$ erg/cm^3•s and $t^* \to \infty$

for Model 2 $Q_o^* = 3.2569 \cdot 10^{-7}$ erg/cm^3•s and $t_o^* = 2219$ Ma,

for Model 3 $Q_o^* = 1.7131 \cdot 10^{-7}$ erg/cm^3•s and $t_o^* = 3248$ Ma.

In the recalculation, a mean density of the lower mantle of $\rho_k = 5.01$ g/cm^3 was used. We assume that the upper surface of the lower mantle is located at a depth of 670 km and the core-mantle boundary at a depth of 2885.3 km. Moreover, the heat flow through the core-mantle boundary is to be so small as to be neglectable. Shock-wave investigations of the electrical conductivity of core-candidate iron alloys demonstrate that the heat flow through the core-mantle boundary will be nearly the same as the average surface heat flow if the geodynamo is thermally driven. Because we believe that the estimations of the order of magnitude of the radiogenic heat source concentrations in crustal and mantle rocks are realistic, the theoretical surface heat flow would exceed considerably the observed average heat flow at the surface. Therefore, we suppose lower temperature gradients in the core and a low heat flow into the mantle. In this case, the geodynamo is compositionally driven.

The heat flow through the upper surface of the lower mantle is to be generally expressed through the Nusselt number N of the convection continuously taking place in the upper mantle. The exponential dependence of viscosity on temperature, which is applicable to both the upper and lower mantle, is essential. The temperature T_2^* on the Earth's surface is to be fixed at 300 K; the temperatures T_1^* and T_o^* on the upper boundary and lower boundary of the lower mantle are functions of the time t^*.

Next, an intermediate consideration follows whose purpose is to express the heat flow density χ_σ^* on the upper boundary of the lower mantle as a function of T_1^* and T_o^*. For convection in the upper mantle, we assume

$$N = c\ R^{1/3} \tag{2}$$

where R is the Rayleigh number

$$R = \frac{g\ \alpha\ \Delta T^*\ h_1^3}{k\ \nu} \tag{3}$$

If we include the oceanic lithosphere in the circular movement of upper-mantle convection, we can assume the upper boundary to be stress-free and the lower boundary to be rigid. The critical Rayleigh number for the rigid-free case is $R_c = 1100.65$. Therefrom and from (2), we obtain c = 0.096 861 6. Turcotte et al. (1973) use the following approximation for the viscosity of the mantle:

$$\eta = 2.76 \cdot 10^3\ T^*\exp\left[(5.222 \cdot 10^4 + 1.087 \cdot 10^{-7} p^*)/T^*\right] \tag{4}$$

where T* is given in K and p* in dyn/cm^2. We simplify the formula for the upper mantle by substituting the mean pressure $p^*_{av} = 1.113 \cdot 10^{11}$ dyn/cm^2 for p* and by substituting the mean temperature $T^*_{av} = (T^*_1 + T^*_2)/2$, which varies with time, for T* in the exponential function. The factor T* in (4) and ΔT^* in (3) are to compensate one another more or less, so that we use the approximation

$$R = R_k\ \exp(-k_4/2T^*_{av}) \tag{5}$$

for the upper mantle, where $k_4 = 2 \cdot 6.432 \cdot 10^4$ K and $R_k = 5.0351 \cdot 10^{20}$ applies. R_k has been determined such that $R = 10^6$ applies now in the upper mantle. McKenzie and Weiss (1975) assume $R = 10^5$ to 10^6 for the upper mantle, but we only require an approximate calculation for our determination of the Nusselt number. For the upper mantle, we have assumed in this recalculation a temperature of 1900 K at a depth of 335 km. The purely diffusive heat flow density at the lower/upper mantle interface is $-\kappa dT^*/dx^*_3$ where κ is the thermal conductivity and x^*_3 the upwards pointing location coordinate. The thermal diffusivity k is $k = \kappa/\rho c_p$. We use

k =	$8 \cdot 10^{-3}$ cm^2/s	for the thermal diffusivity,
α =	$2 \cdot 10^{-5}$ K^{-1}	for the coefficient of thermal expansion,
c_p =	$1.2 \cdot 10^7$ erg/g·K	for the specific heat at constant pressure.

Hence, the entire heat flow density at the upper boundary of the lower mantle can be expressed by

$$\chi^*_\sigma = k\rho\ c_p\ \frac{T^*_1 - T^*_2}{h_1}\ N \tag{6}$$

where h_1 is the thickness of the upper mantle. Therefrom and from (2) and (5) follows

$$\chi_\sigma^* = k\rho\, c_p\, c\, \frac{T_1^* - T_2^*}{h_1} \left[R_k \exp\left[-k_4/(T_1^* + T_2^*)\right]\right]^{1/3} \tag{7}$$

We conclude from the movement of the lithospheric plates that convection is continuously going on in the upper mantle. Therefore, the heat flow on the upper boundary of the lower mantle is generally defined by (7). The low heat production density of the lower mantle is insufficient for continuously driving convection there. But since, at the same time, the lattice and radiative heat conductivities are too low for dissipating the heat from the lower mantle (Pitt and Tozer, 1970), the temperature there is slowly rising. Due to the temperature dependence of viscosity, viscosity decreases and the Rayleigh number increases until the critical Rayleigh number is exceeded also in the lower mantle. This gives rise to convection in the lower mantle, which promotes convection in the upper mantle and the magmatic and tectonic activity in the continental lithosphere and also influences the magnetic field of the Earth. Lower-mantle convection based on this mechanism must soon die down again, because the temperature decreases as a result of the heat dissipation, so that the Rayleigh number again falls beyond the critical value, i.e., the energy source is not sufficient for *continuously* driving the motor.

Strictly speaking, one would have to solve the following set of equations for the lower mantle

$$\rho\left(\frac{\partial \vec{v}^*}{\partial t} + \vec{v}^* \cdot \nabla^* \vec{v}^*\right) = -\rho g \vec{k} - \nabla^* p^* + \tag{8}$$

$$+ k_1 \frac{\partial}{\partial x_k^*} \left[T^* e^{k_2 T_m(p^*)/T^*} \left(\frac{\partial v_i^*}{\partial x_k^*} + \frac{\partial v_k^*}{\partial x_i^*}\right)\right],$$

$$\nabla^* \cdot \vec{v}^* = 0, \tag{9}$$

$$\frac{\partial T^*}{\partial t^*} + \vec{v}^* \cdot \nabla^* T^* = k\, \nabla^{*2}\, T^* + Q^*(t^*, x_3^*)/\rho_k\, c_p, \tag{10}$$

$$\rho = \rho_k \left[1 - \alpha (T^* - T_k)\right]. \tag{11}$$

Variables marked by * are to denote dimensional variables which are subsequently replaced by dimensionless variables, with the asterisk being deleted. In addition to the variables already introduced we have: ρ = density, $\vec{v}^* = v_k^*$ = velocity vector, $\vec{r}^* = x_k^*$ = location vector, $\vec{k}$ = upwards pointing unit vector, g = amount of gravitational acceleration; k_1, k_2, ρ_k, T_k are constants, h_o is the thickness of the lower mantle. The boundary conditions, which have already been specified with respect to the heat flow densities, have to be complemented by

$$v_3 = \frac{\partial v_1}{\partial x_3} = \frac{\partial v_2}{\partial x_3} = \frac{\partial^2 v_3}{\partial x_3^2} = 0 \quad \text{for} \quad x_3 = 0 \quad \text{and} \quad x_3 = h_o \tag{12}$$

because the lower mantle, by virtue of its high viscosity, compared with the upper mantle and the outer core, high viscosity, can be assumed to have quasi-stress-free boundary faces. The upwards directed component x_3 of the location vector in its dimensionless form is denoted by z; the zero-point is to be located in the core-mantle boundary. Now we introduce dimensionless variables.

$$t^* = t\,\frac{h_o^2}{k}; \quad \vec{r}^* = \vec{r} h_o; \quad T^* = T\,\frac{kk_1}{g\,\alpha\,h_o^3\,\rho_k}; \tag{13}$$

$$p^* = p\,\frac{kk_1}{h_o^2}; \quad \vec{v}^* = \vec{v}\,\frac{k}{h_o}; \quad \frac{g\,\alpha\,Q^*(t^*, x_3^*)h_0^5}{k^2 k_1 c_p} = Q(t);$$

Next, we shall each time consider the static case up to the commencement of the change, i.e., of the convection episode. From (10) we then obtain in dimensionless variables the following partial differential equation

$$\frac{\partial T}{\partial t} - \frac{\partial^2 T}{\partial z^2} = Q(t, z) \tag{14}$$

Q(t, z) is a given function.

According to Model 1, we have $Q = 6.445\,96 \cdot 10^{11}$,
according to Model 2, we have $Q_o = 4.897\,8 \cdot 10^{12}$,
according to Model 3, we have $Q_o = 2.576\,2 \cdot 10^{12}$,

(1) with (13) having to be taken into account for the last two models. t = 0 designates the moment at which the development commenced. In the model, it is equalled to the accretion of the Earth. t_σ designates the onset of the σ-th convection episode. At first, the variation of the temperature T(z, t) in the interval $0 < z < 1$ is calculated for the first convectionless interval $0 < t \leqq t_1$. The end of the interval t_1 (or, generally t_σ) is determined by the surpassing of the critical Rayleigh number in the lower mantle, i.e., by the development of the temperature behaviour during the calculation, as will be discussed in greater detail below. In the calculation, the initial conditions

$$T(z, O) = 0 \quad \text{in} \quad 0 < z < 1 \tag{15}$$

and the boundary conditions

$$\frac{\partial T(0,t)}{\partial z} = \psi \quad \text{in} \quad 0 < t < t_1, \tag{16}$$

$$\frac{\partial T(1,t)}{\partial z} = \chi_\sigma\,[T(1,t)] \quad \text{in} \quad 0 < t < t_1 \tag{17}$$

have to be satisfied. In the calculations, we used $\psi \equiv 0$, but $\psi = \text{const} \neq 0$ can be easily used later on if one wishes to study the effects of a heat flow from the core. The boundary condition (17) is non-linear, because from (7) follows

$$\chi_\sigma = k_7\,[T_2 - T_u^{(\sigma)} - T(1,t)] \;\cdot \tag{18}$$
$$\cdot\;[\exp[-k_8/(T_2 + T_u^{(\sigma)} + T(1,t))]]^{1/3}\;.$$

The values of the constants are $k_7 = 2.547\,7371 \cdot 10^6$ and $k_8 = 1.8959 \cdot 10^{13}$. The quantity $T_u^{(\sigma)}$ is a dimensionless temperature which is newly calculated according to the results of the preceding processes for each $\sigma > 1$, i.e., for each convectionless interval after the first convection episode, while $T_u^{(1)}$ is presupposed as a constant initial temperature.

The set of equations (14) to (18) can be replaced by (18) to (21), with the last equation being a non-linear Hammerstein integral equation which facilitates the numerical calculation:

$$T(z,t) = \int_0^t \int_0^1 G(z,\zeta,t-\tau)Q(\tau,\zeta)d\zeta d\tau \;- \tag{19}$$
$$-\;\psi\int_0^t K(z,t-\tau)d\tau + \int_0^t K(1-z,t-\tau)\chi_\sigma(T_1(\tau))d\tau;$$

$$T_0(t) = \int_0^t \int_0^1 K(\zeta,t-\tau)Q(\tau,\zeta)d\zeta d\tau \;- \tag{20}$$
$$-\;\psi\int_0^t K(0,t-\tau)d\tau + \int_0^t K(1,t-\tau)\chi_\sigma(T_1(\tau))d\tau;$$

$$T_1(t) = \int_0^t \int_0^1 K(1-\zeta,t-\tau)Q(\tau,\zeta)d\zeta d\tau \;- \tag{21}$$
$$-\;\psi\int_0^t K(1,t-\tau)d\tau + \int_0^t K(0,t-\tau)\chi_\sigma(T_1(\tau))d\tau$$

where

$$T_0(t) \equiv T(0,t) \quad \text{and} \quad T_1(t) \equiv T(1,t). \tag{22}$$

G denotes the Green function

$$G = G(z,\zeta,t) = \tfrac{1}{2}[\vartheta(\frac{z-\zeta}{2}, L) + \vartheta(\frac{z+\zeta}{2}, L)] \tag{23}$$

where ϑ denotes the Jacobi theta function

$$\vartheta(s, L) = 1 + 2 \sum_{n=1}^{\infty} L^{n^2} \cos 2n\pi s \tag{24}$$

in $0 \leqq s \leqq 1$ and $0 \leqq L < 1$.

$$L = L(t) = \exp(-\pi^2 t). \tag{25}$$

The kernel is given by

$$K(z, t) = \vartheta(s, L) \tag{26}$$

with $s = z/2$. Our main problem now is to solve (21). Once we have found $T_1(t)$, we can determine $T(z, t)$ and $T_o(t)$ with the help of (19) and (20). T_1, T and T_o were calculated for equidistant steps Δt. After each step, the Rayleigh number R in which the temperature dependence of the viscosity of the lower mantle was taken into account was determined:

$$R = \beta \; \overline{\exp[k_{15}(z)/(T(z) + T_u^{(\sigma)})]} \tag{27}$$

where $\beta = \overline{|dT/dz|}$. Cross bars signify averaging for the interval $0 \leqq z \leqq 1$, i.e., for the entire lower mantle. k_{15} is determined as outlined in the previous section and according to the melting temperature curve T_m^* by Stacey (1977). We now assume that for a radius of 4600 km (i.e., about in the middle of our layer) we have $T_m^* = 2776$ K, $\eta = 3 \cdot 10^{25}$ Pa•s and $k_{15}(0.5) = -3.5441 \cdot 10^{12}$. For each step, a check is made whether

$$R \geqq R_c \tag{28}$$

has been reached, i.e., whether convection has already started. $R_c = 657.5$ has been used because both the upper and lower boundaries of the lower mantle are almost stress-free. Then, thermal compensation through convection is to take place. The initial temperature of the next convectionless interval is calculated by

$$T_u^{(\sigma+1)} = \int_0^1 T(z, t_\sigma)dz + T_u^{(\sigma)} \tag{29}$$

i.e., the mean value is taken. From interval to interval, the index σ is increasing by one.

5. VERIFICATION

Table I shows the most important numerically calculated parameters of the three models developed by Walzer and Maaz (1983). The question now is whether these theoretically gained results can be verified by values obtained in observations. Gastil (1960) has found that the numerous radioactive age determinations that have been mainly carried out on granites are by no means uniformly distributed or just slightly scattered over the time axis. There exists pronounced maxima at irregular intervals (see Fig. 3). This surging and fading of magmatism is a much slower process than the episodes which, as synorogenic magmatism, are connected with Stille's disputed orogenetic phases. The dispute going on among geologists with respect to this significantly shorter phenomenon does not affect the Gastil curve. This curve proves to be, also, by the way, if further age determinations are included (Kölbel, 1971), independent of the continent investigated. Thus, it cannot be explained in terms of local processes (e.g., in the lithosphere), but we

TABLE I. SOME COMPUTER RESULTS ACCORDING TO WALZER AND MAAZ (1983)

Model	$T_u^{(1)*}$/K	σ	1	2	3	4	5
		$T_1^*(t_\sigma)$/K	1879.279	1902.670	1920.338	1932.095	1940.885
1	1841	t_σ^*/Ma	1767.065	2802.250	3585.310	4109.274	4502.174
		$(t_\sigma^*-t_{\sigma-1}^*)$/Ma	1767.065	1035.186	783.060	523.964	392.900
		$T_1^*(t_\sigma)$/K	1893.068	1922.992	1936.561	1945.026	
2	1618	t_σ^*/Ma	2894.117	3652.610	4108.232	4449.622	
		$(t_\sigma^*-t_{\sigma-1}^*)$/Ma	2894.117	758.493	455.622	341.390	
		$T_1^*(t_\sigma)$/K	1893.153	1920.408	1933.917	1942.646	
3	1725	t_σ^*/Ma	2819.967	3633.377	4128.443	4496.241	
		$(t_\sigma^*-t_{\sigma-1}^*)$/Ma	2819.967	813.410	495.067	367.797	

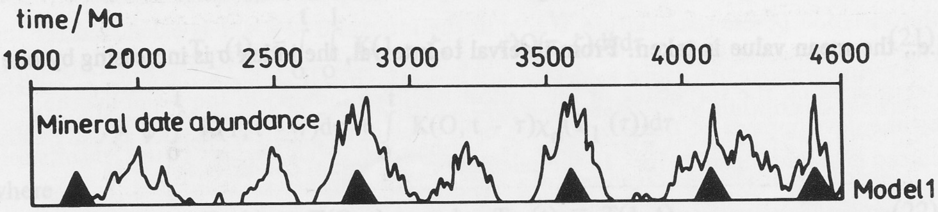

Fig. 3: Comparison of the global abundance of the radiometric age determinations according to Gastil (1960) with the convection episodes of Model 1 according to Walzer and Maaz (1983).

are dealing with a global phenomenon here. The idea suggests itself that the curve is also defining the rising and falling of the mean heat flow radiated into outer space, and this is the decisive variable in the thermal and tectonic history of the Earth. For these reasons, the underlying mechanism must involve large parts of the Earth and must be of primary importance also from the energetic point of view. Since the lower mantle is that part of the Earth which has the greatest mass and which, because viscosity is increasing downwards, is probably not always flowing convectively, it seems justified to look for a feedback mechanism there, which provides an explanation for the observed slow rise and fall in the global magmatic activity. In contrast to other mechanisms that have been suggested, the mechanism calculated here shows for the first time a *quantitative* agreement with the most important maxima of the Gastil curve. In Figures 3 to 5, a comparison is made between the convection episodes (overturns) according to the Models 1 to 3 and the Gastil curve. There is one unrealistic aspect in Model 1: In this model, the heat production density is constant. Thus, it is not astonishing that the first turn (see Fig. 3) does not have a corresponding maximum in the granite frequency.

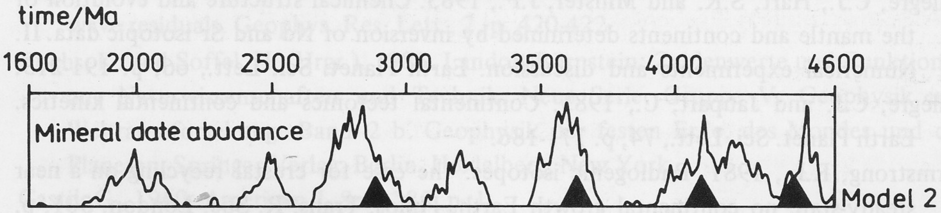

Fig. 4: Comparison of the global abundance of the radiometric age determinations according to Gastil (1960) with the convection episodes of Model 2 according to Walzer and Maaz (1983).

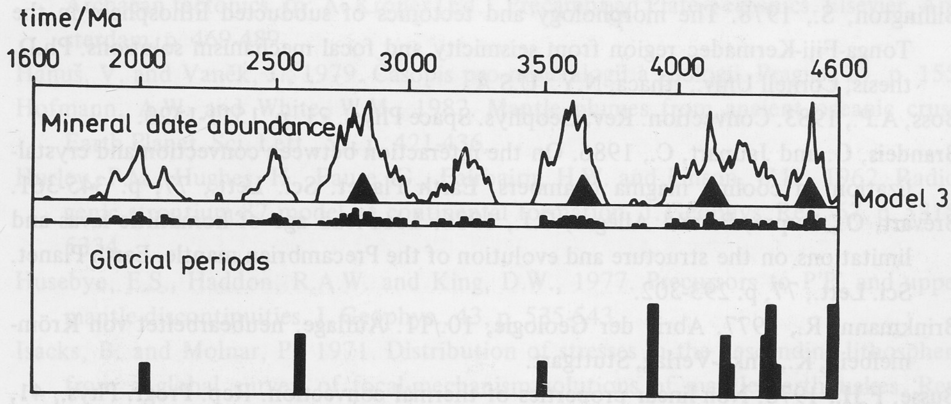

Fig. 5: Comparison of the global abundance of the radiometric age determinations according to Gastil (1960), the glacial periods according to Brinkmann (1977) and the convection episodes of Model 3 according to Walzer and Maaz (1983). This model probably is the best one, also with respect to the assumptions as to the heat source distribution.

Model 2 (Fig. 4) is more realistic, but the numbers assumed there appear to be still too far away from the truth. Model 3 (Fig. 5) is the best one. Here, each of the 4 greatest maxima in the granite frequency curve exactly corresponds to a convection episode of the lower mantle and, with respect to time, these episodes are precisely located at the correct place. This result is all the more remarkable since the intervals between the overturns are not identical. The smaller maxima are an indication of an additional mechanism with lesser energy budget. This success makes the author hopeful that, with Model 3, an essential feature in the development of the Earth has been found.

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