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TEMPERATURE GRADIENT AND THE EVOLUTION OF THE EARTH'S MANTLE

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Conditions in the interior of the Earth are determined by two parameters P and T. Pressure P is known with sufficient accuracy as it follows from comparison of various Earth models. Unfortunately the temperature T in the interior of the Earth is determined with considerably lower certainty. However, the Earth can be considered as a heat engine and the action of this engine depends strongly on its temperature conditions. A knowledge of T, $\partial T/\partial z$ and $\partial T/\partial t$ is of extreme importance for the development of geophysical theories concerning the problem of the evolution of our planet.

Calculations of the Earth's temperature based on the thermal history of the planet can give only very uncertain results depending on conditions of formation and evolution of the Earth (MACDONALD, 1959; LUBIMOVA, 1969).

Determination of the electrical conductivity in the interior of the Earth can give us valuable information concerning the temperature at various depths. Unfortunately, electrical conductivity of rocks is not only sensitive to temperature but strongly depends on impurities and other uncertain properties of the material under consideration.

Seismological and volcanological data have been used in geothermics for a long time. But only the recent achievements in high-pressure experiments permit us to use these sources of information with some reliance.

Apparently, knowing $\partial T/\partial z$ at various depths one can calculate T in the interior of the Earth as well as the heat flow q at respective points, the thermal conductivity K being known.

1. The thermal gradient at the surface of the Earth

The mean value of $(\partial T/\partial z)_0$ for continental and oceanic parts of the Earth's surface can be obtained from the measured values of q_0 using the formula

$$q_0 = K_0 \left(\frac{\partial T}{\partial z}\right)_0, \tag{1}$$

where K_0 is the heat conductivity at the surface. For that purpose the most suitable values of q are those obtained from $5° \times 5°$ averages (VON HERZEN and LEE, 1969). These averages are free from irregular fluctuations and give a mean terrestrial heat flow not strongly dependent on the distribution of individual measurements as compared with the simple mean of all measured values.

Both for oceans and continents

$$q_0 = 1.3 \times 10^{-6} \text{ cal cm}^{-2}\text{ s}^{-1}$$

To obtain K_0 data for typical rocks were employed (CLARK, 1966). The adopted values were:

$$K_0 = 7.1 \times 10^{-3} \text{ cal cm}^{-1}\text{ s}^{-1}\text{ °C}^{-1}$$

for continental crust, and

$$K_0 = 5.2 \times 10^{-3} \text{ cal cm}^{-1}\text{ s}^{-1}\text{ °C}^{-1}$$

for oceanic crust. Hence $(\partial T/\partial z)_0$ was found to be 18 °C/km for continents and 25 °C/km for oceans.

2. The thermal gradient at the depth of magma sources

Primordial sources of magmas in continental regions of the Earth are located, according to seismic investigations, at a depth of 100–150 km (BLOT, 1964; FEDOTOV, 1968; FEDOTOV and SLAVINA, 1968). For oceanic regions the corresponding depth is 60–80 km (POWERS, 1955; EATON, 1962), that is, the depth of the low-velocity layer (ANDERSON, 1967; PRESS, 1968). Also, the nature of the low-velocity layer is not quite certain, but there are strong evidences in favour of partial melting as the principal cause of its origin (BIRCH, 1969; MAGNITSKY and ZHARKOV, 1969; TAKEUCHI *et al.*, 1968). Results of high-pressure experiments in the presence of H_2O support this point of view (LAMBERT and WYLLIE, 1969; SCLAR, 1969).

To estimate the temperature at which melting begins at a given depth, it is necessary to know the pressure and the composition of rocks. For the uppermost parts of the mantle the accepted composition is that of garnet peridotites (garnet–lherzolite) (O'HARA, 1968) maybe with an admixture of eclogites (SOBOLEV, 1968).

Keeping in mind the high content of H_2O vapour in extruding lavas, one must use results of melting experiments in the presence of H_2O. Experiments with dry rocks would give exaggerated temperatures. Recently obtained data on melting in the presence of H_2O for lherzolite inclusions in basalts give the following results (KUSHIRO *et al.*, 1968): The temperature of the solidus at a depth of 100 km (if $P_{\text{total}} = P_{H_2O}$) must be near 1000 °C; for fusion of 10% of the material the temperature ought to be 100 °C higher. Hence for a depth of 100 km one must adopt a temperature

$$T_{100} = 1100\ ^\circ\mathrm{C},$$

and if $P_{H_2O} < P_{\text{total}}$,

$$T_{100} \approx 1200\ ^\circ\mathrm{C}.$$

Consequently,

$$\left(\frac{\partial T}{\partial z}\right)_m = 11.5\ ^\circ\mathrm{C/km}.$$

Keeping in mind that $(\partial T/\partial z)_0 = 18$ °C/km, for a depth of 100 km we get

$$\left(\frac{\partial T}{\partial z}\right)_{100} = 5\ ^\circ\mathrm{C/km}.$$

The same procedure gives for the oceanic parts of the Earth:

$$\left(\frac{\partial T}{\partial z}\right)_{60} = 8\ ^\circ\mathrm{C/km}.$$

Adopting the hypothesis that the formation of a low-velocity layer is caused by the influence of high-temperature gradients, it was found by THOMSON (1967) that $(\partial T/\partial z)_{\min}$ at corresponding depths is 4.8 °C/km under the continents and 9.5 °C/km under the oceans. This is in fair agreement with our estimations.

3. The thermal gradient at the top of the transition layer

The results of high-pressure experiments leave no doubt that the transition layer C is caused mainly by some kind of phase transformation in rocks. In accordance with the latest seismological informations (JONSON, 1967; D. L. ANDERSON, 1967; PRESS, 1968), layer C is divided into sublayers. The first sublayer with high velocity and density gradient due to phase transitions was discovered at a depth of about 400 km, its thickness being in the range of 50–80 km.

The principal component of the upper mantle is olivine. Phase transformations in solid solutions of $(Mg, Fe)_2SiO_4$ have been comprehensively investigated. Hence it is possible to calculate the transition in the P–T-plane using routine procedures (MAGNITSKY, 1969).

The transition from the olivine to the spinel phase is complicated due to the splitting of the transition at the point $(Mg_{0.8}, Fe_{0.2})_2\ SiO_4$ into two branches. Nevertheless, approximate calculations can be done neglecting the difference between the spinel ($\rho = 3.54$ g/cm^3) and distorted spinel ($\rho = 3.48$ g/cm^3) phases (RINGWOOD, 1968; FUJISAWA, 1968). Boundaries of the transition band were obtained as follows:

$$T_1 = -1289 + 16.75P,$$
$$T_2 = -787 + 15.75P,$$

where T is in °C and P in kb.

Adopting from seismological data the pressure at the upper and lower boundaries of the transition zone (123 and 151 kb) one obtains the corresponding value of the temperature gradient $\partial T/\partial z$. For various models, gradients were obtained in the range 1–2 °C/km. As an average value was adopted

$$\left(\frac{\partial T}{\partial z}\right)_{400} = 1.3\ ^\circ\mathrm{C/km}.$$

In these calculations data were employed from FUJISAWA (1968), RINGWOOD (1968), RINGWOOD and MAJOR (1969), and KAWAI *et al.* (1969).

4. The thermal gradient in the uppermost part of the lower mantle

The lower mantle can be regarded as a sufficiently homogeneous layer, such that it is possible to get $\partial\phi/\partial P$ and $\partial\phi/\partial z$, where $\phi = K/\rho = v_P^2 - \frac{4}{3}v_S^2$.

Employing the "law of corresponding states" (O. L. ANDERSON, 1966) values were calculated of $\partial\phi/\partial P$ and $\partial\phi/\partial T$ at a depth of 1200 km, and finally the value of $\partial T/\partial z$ (MAGNITSKY, 1968). The results obtained were:

$$\left(\frac{\partial T}{\partial z}\right)_{1200} = 1.8\text{–}1.2\ ^\circ\text{C/km}.$$

Mean value: 1.5 °C/km.

It is necessary to emphasize that all thermal gradients obtained exceed the adiabatic gradients at corresponding depths: that is undoubtedly evidence against the hypothesis of thermal convection in the mantle. The temperatures in the mantle according to the obtained gradients are as in table 1.

TABLE 1

Depth (km)	T (°C)
100	1100
400	2100
1200	3200

It seems to be very important that within the limits of accuracy the thermal gradients at 400 and 1200 km have the same value of 1.5 °C/km. Supposing this value to be valid throughout the whole lower mantle, one obtains the temperature at the boundary of the Earth's core to be 5500 °C, that is, near the melting point. It must be emphasized that the thermal gradient in the mantle is to the order of magnitude the same as that of the melting curve.

To estimate the heat flux at various dephts it is necessary to have values of the thermal conductivity K at corresponding points. At a depth of about 100 km, with the assumption of a mainly olivine composition it is found that

$$K = 0.009\ \text{cal cm}^{-1}\ \text{s}^{-1}\ ^\circ\text{C}^{-1}$$

(FUJISAWA *et al.*, 1968). For conditions existing at depths of 400 and 1200 km there are no experimental data. Calculations give

$$K = 0.05\ \text{cal cm}^{-1}\ \text{s}^{-1}\ ^\circ\text{C}^{-1}$$

(LUBIMOVA, 1968). Hence for the values of the heat flux q we obtain those listed in table 2.

The difference

$$q_0 - q_{100} = 0.8 \times 10^{-6}\ \text{cal cm}^{-2}\ \text{s}^{-1}$$

seems to be quite reasonable, keeping in mind that 10 km of granite generate an increment of the heat flow of 0.6×10^{-6} cal cm^{-2} s^{-1}, and 10 km of basalt of 0.2×10^{-6} cal cm^{-2} s^{-1}. The difference

TABLE 2

Depth (km)	q (10^{-6} cal cm^{-2} s^{-1})
0	1.3
100	0.5
400	0.7
1200	0.7

$$q_{100} - q_{400} = -0.2 \times 10^{-6}\ \text{cal cm}^{-2}\ \text{s}^{-1}$$

is peculiar. Taking into consideration that 300 km of chondritic material or of peridotites give 0.1–0.2 × 10^{-6} cal cm^{-2} s^{-1}, we must explain the loss in heat flux of 0.4×10^{-6} cal cm^{-2} s^{-1}. Assuming that this loss is due to heating and to the phase transformation spinel–olivine, and adopting a value of heat of transformation of 45 cal/g, one obtaines for the rate of displacement of the phase boundary

$$\frac{\partial z}{\partial t} \approx 0.1\ \text{cm/y}.$$

The corresponding increase of volume is 10%. Hence it is reasonable to expect an increase of the Earth's radius of

$$\frac{\partial R}{\partial t} \approx 0.01\ \text{cm/y}$$

or 1 km per 10^7 y.

Thus the investigation of thermal gradients in the Earth's interior suggests the expansion of the circumference of the Earth's surface to be of the order of 60 km per 10^8 y. That seems to be sufficient to explain the creation of the World Rift system. These conclusions are of a quite preliminary character due to the low accuracy of the data used.

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