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Geothermal Gradients and Temperatures in the Mantle and the Problem of Fusion

V. A. Magnitsky

Institute of the Physics of the Earth, Academy of Sciences
Moscow, USSR

The existence of the low-velocity layer (or layers) in the uppermost parts of the mantle, at least in oceanic and tectogenic regions, is now accepted without any doubt. Different explanations for the origin of the low-velocity layers have been suggested. One suggestion has been that partial melting is the principal cause; however, this suggestion presents two problems: (1) poor knowledge of the temperature at each respective depth, and (2) discovery in some areas of high-velocity layers that are situated just above the low-velocity zones. These two problems are discussed in this paper.

The existence of the low-velocity layer (or layers) in the uppermost parts of the mantle, at least in oceanic and tectogenic regions, is now accepted without any doubt [*Press*, 1970; *Hales et al.*, 1968; *Anderson*, 1967]. There is strong proof in favor of the presence of a shear-wave low-velocity zone even in shield areas [*Dorman*, 1969], in spite of a probable absence of *P*-wave channels in these regions. In various parts of the earth low-velocity layers have different thicknesses and are situated at different depths.

The problem of the physical or physico-chemical nature of low-velocity layers is one of the most important in geophysics, geotectonics, and magmatism, because these layers are commonly identified with the zone of the asthenosphere. Hence the problem of the origin of low-velocity layers is intimately connected with problems of isostatic readjustment due to low viscosity in these layers [*Artyushkov*, 1966] and with problems of partial melting. Consequently the whole problem of the thermics of the earth's interior is involved in the consideration of the nature of the low-velocity layer.

Many authors have suggested different explanations for the origin of low-velocity layers. For review see *Birch* [1969] and *Magnitsky and Zharkov* [1969]. Examining various aspects of the problems, many scientists concede the suggestion of partial melting as the principal cause of low-velocity zones [*Press*, 1970; *Anderson and Sammis*, 1970; *Takeuchi et al.*, 1968; *Fedotov*, 1966]. However, this suggestion confronts two difficulties: (1) poor knowledge of the temperature at each respective depth, and (2) discovery in some areas of high-velocity layers that are situated just above the low-velocity zones. These two items are the subject of this paper.

Determining Geotherms

The principal method of calculating geotherms is based on solving the steady-state thermal problem. To solve this problem, the heat flux at the earth's surface, the mechanism of heat transfer, the effective thermal conductivity and heat capacity, and their dependence on temperature, pressure, and composition of the medium must be known. In addition, distribution of heat sources in the crust and mantle must be known. In a non-steady problem, i.e., thermal history, the difficulties only augment the investigation [*Lubimova*, 1968]. Unfortunately, the present distribution of heat sources in the mantle and crust of the earth is not known with sufficient plausibility. Our information on this subject is too meager. Hence geotherms calculated in such a manner are very uncertain. It would be highly desirable to employ a different method of calculating geotherms to check the results obtained with the classical method.

Here the adopted method is based on calculation of geothermal gradients at various depths in the crust and mantle. The gradient at the surface is calculated at a chosen region without difficulty if the heat flux and thermal conductivity are known. Calculations were performed for the western part of the Black Sea and a typical ocean region. With proper data [*Lubimova*, 1968; *Simmons and Horai*, 1968; *Clark*, 1966] the following values were calculated:

$(dT/dZ)_0 = 18°\text{C/km}$ Black Sea

$(dT/dZ)_0 = 25°\text{C/km}$ typical ocean bottom

It is possible to calculate dT/dZ at any depth by using the ordinary formula

$$\frac{dv}{dZ} = \left(\frac{\partial v}{\partial p}\right)_{T,C,\phi} \rho g + \left(\frac{\partial v}{\partial T}\right)_{p,C,\phi} \frac{dT}{dZ} + \left(\frac{\partial v}{\partial C}\right)_{p,T,\phi} \frac{dC}{dZ} + \left(\frac{\partial v}{\partial \phi}\right)_{p,T,C} \frac{d\phi}{dZ} \qquad (1)$$

where v is the velocity of seismic waves, p is the pressure, ρ is the density, g is the acceleration of gravity, C is the composition factor, and ϕ is the phase transformation factor.

All needed partial derivatives, ρ, g, and the dependence of composition and phase transition on Z can be determined. The term dv/dZ is supposed to be available from seismic data. Unfortunately dv/dZ as a rule is unknown. Thus (1) could not be used, except for a very rough appraisal in some instances.

Seismic sounding in the Black Sea with sea-bottom seismographs offered an exceptional opportunity to employ (1) for determining dT/dZ. Travel-time curves of head waves were obtained at the Mohorovicic discontinuity (depth, 18 km; v_p, 8.2 km/sec) and at another discontinuity (depth, 30 km; v_p, 8.8 km/sec), and the amplitudes of these waves and of refracted waves penetrating the underlying layer were observed, as is shown in Figure 1. From Figure 1 it can be seen that dv/dZ at a depth of 20–30 km is 0.017 sec^{-1} and that just below 30 km, dv/dZ is 0.015 sec^{-1}.

There are no grounds to suspect any phase transition between the discontinuities at 18 and 30 km just below 30 km. Hence in (1) the last term could be omitted. There are also no reasons to suspect considerable variations in composition, and so, as a first approximation, $dC/dZ = 0$.

In the two remaining right-hand terms in (1), ρ and g are well known. The problem is to obtain $(\partial v/\partial p)_{T,C,\phi}$ and $(\partial v/\partial T)_{p,C,\phi}$. The value of $(\partial v/\partial T)_{p,C,\phi}$ was accepted as -4×10^{-4} km/sec °C [*Anderson et al.*, 1968] as typical for rocks of the uppermost mantle. The difficulty in estimating $(\partial v/\partial p)$ arose from the experimental fact of the possible effect of pores at the pressures under consideration. Thus the value

$$(\partial v/\partial p)_{T,C,\phi} = 0.065 \text{ km/sec kb}$$

which was obtained for dunite and eclogite nodules and peridotites and pyroxenites from Kola peninsula [*Maghnani and Woollard*, 1968; *Volarovich and Levykin*, 1965], was accepted as suitable. By using all these values and (1), the value of dT/dZ at $Z = 30$ km at the Black Sea area was determined as $(dT/dZ)_{30} \approx 10°\text{C/km}$.

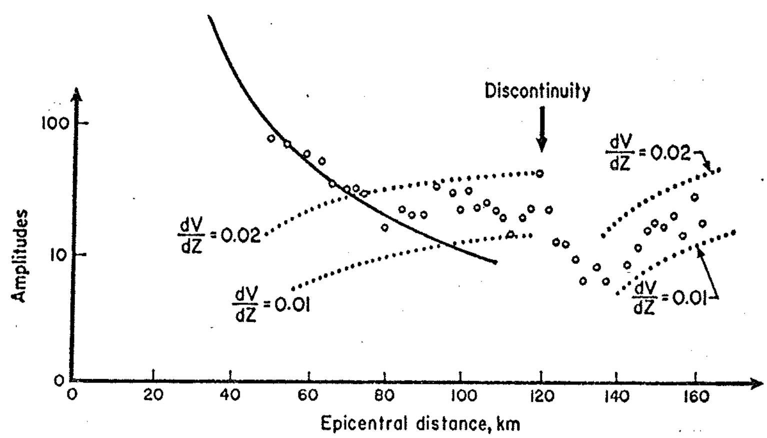

Fig. 1. Theoretical and observed amplitudes of P waves in the Black Sea region. The solid line represents the theoretical amplitudes (in millimicrons) of head waves at the Mohorovicic discontinuity. The dotted line represents the theoretical amplitudes of penetrating waves calculated with various values of dv/dZ. Open circles are the observed amplitudes.

An average of dT/dZ in the low-vel could be estimated in the manner des low.

If a low-velocity layer in the upperm tle at the depth 60–200 km is the av of partial melting, dT/dZ must be o of the gradient of the respective soli Unfortunately the composition of t most mantle is still not known exac are different opinions in regard to lem [*Clark and Ringwood*, 1964; 1968; *O'Hara*, 1968]; yet the prob entirely hopeless. Most silicates at corresponding to a depth of ~100 melting gradients in the range 2 Even the presence of H_2O does n cantly alter these values [*Kushiro et Lambert and Wyllie*, 1970; *Kadik a rov*, 1970].

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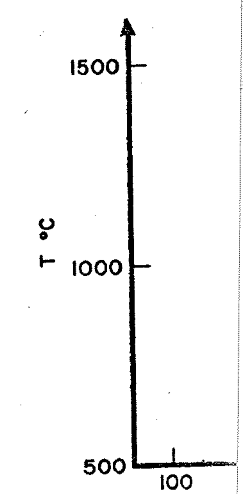

Fig. 2. Determination of tem are the limits of the olivine–spi transition layer, and b and c ar km, respectively.

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$$(\partial v/\partial p)_{T,c,\phi} = 0.065 \text{ km/sec kb}$$

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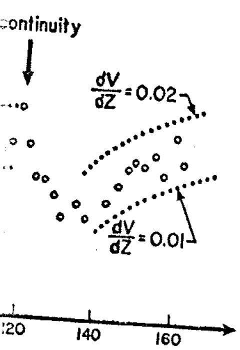

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...n average of dT/dZ in the low-velocity zone could be estimated in the manner described below.

If a low-velocity layer in the uppermost mantle at the depth 60–200 km is the average zone of partial melting, dT/dZ must be of the order of the gradient of the respective solidus curve. Unfortunately the composition of the uppermost mantle is still not known exactly. There are different opinions in regard to this problem [*Clark and Ringwood*, 1964; *Sobolev*, 1968; *O'Hara*, 1968]; yet the problem is not entirely hopeless. Most silicates at pressures corresponding to a depth of ~100 km have melting gradients in the range 2°–5°C/km. Even the presence of H_2O does not significantly alter these values [*Kushiro et al.*, 1968; *Lambert and Wyllie*, 1970; *Kadik and Khitarov*, 1970].

Another widespread point of view is that the low-velocity zone is caused by the prevailing effect of the temperature-gradient term in (1). In this way, and for various compositions of the upper mantle, the critical temperature gradients were obtained for S waves in the range 3.2°–5.5°C/km and for P waves in the range 5.5°–7.1°C/km [*Birch*, 1969].

Hence in the low-velocity zone the value $(dT/dZ)_{60\text{-}200} = 3°$–5°C/km could be accepted.

At present there is little doubt that the rapid velocity increase near the depth of 400 km is associated with olivine–spinel transformation [*Press*, 1970; *Ringwood*, 1970; *Fujisawa*, 1968]. Consequently it is possible to estimate the value of dT/dZ at the depth of ~400 km. High-pressure experiments give a phase diagram in the Mg_2SiO_4-Fe_2SiO_4 system [*Akimoto and Fujisawa*, 1968; *Ringwood and Major*, 1970; *Kawai et al.*, 1970]. The composition of olivine was adopted as 90% Mg_2SiO_4. The slope of the transition curve was accepted as 62 bars/°C [*Akimoto and Fujisawa*, 1968]. Nearly similar results were obtained by using a routine procedure given by *Magnitsky and Kalashnikova* [1970]. Boundaries of the olivine–spinel transformation zone are plotted in Figure 2. The thickness of the seismic transition zone was taken in the range 50–80 km [*Johnson*, 1967; *Green and Hales*, 1968; *Press*, 1970]. The intersection of these boundaries with the

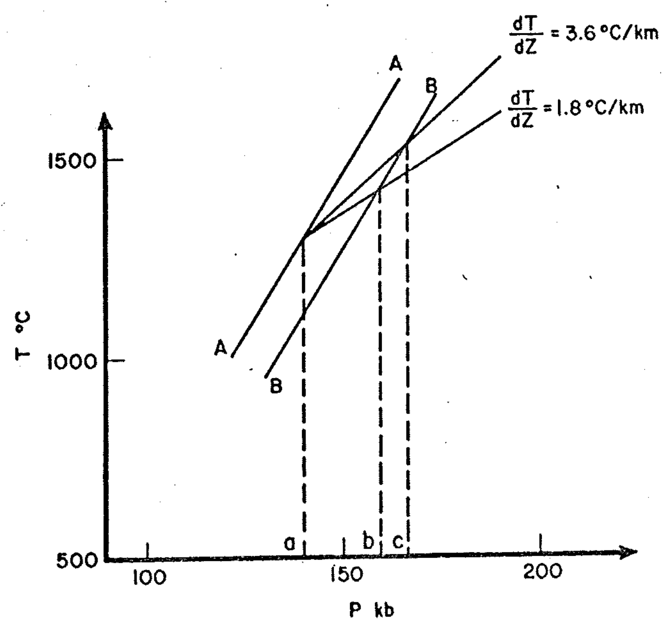

Fig. 2. Determination of temperature gradient in the transitional layer. *A–A* and *B–B* are the limits of the olivine-spinel transformation zone; *a* is the upper boundary of seismic transition layer, and *b* and *c* are lower boundaries corresponding to thickness of 50 and 80 km, respectively.

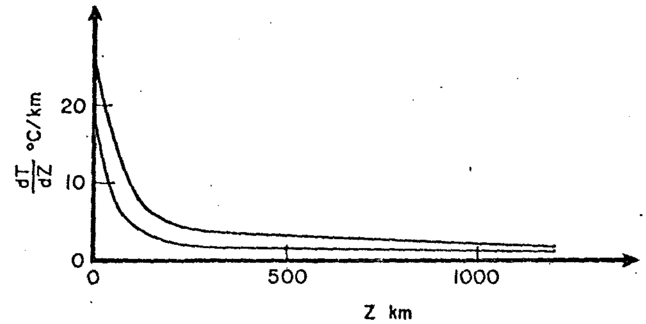

Fig. 3. Limits of dT/dZ in the mantle.

olivine–spinel transition zone gives a value $(dT/dZ)_{400} = 1.8°–3.6°C/km$, as can be seen from Figure 2. Finally dT/dZ could be calculated at the depth of ~1200 km by using the law of corresponding states. This method gives $(dT/dZ)_{1200} = 1.2°–1.8°C/km$ [*Magnitsky*, 1968].

Plotting all values of dT/dZ obtained above, we draw, in Figure 3, two curves representing, as a first approximation, the upper and lower boundaries of geothermal gradients. Integration of these curves gives upper and lower limits of expected temperature in the mantle. In Figure 4 these temperature limits and, for comparison, continental and oceanic geotherms according to *Clark and Ringwood* [1964] are plotted. Comparison with solidi of some rocks in the presence of H_2O are also plotted in Figure 4. *Kushiro et al.* [1968] and *Lambert and Wyllie* [1970] strongly support the idea that partial melting is the principal cause of the low-velocity layer.

High-Velocity Layer

Another problem is how to explain the existence of the layer with high-velocity *P* waves just above the low-velocity zone. Layers of this kind were discovered in the Black Sea region, under some parts of Indian Ocean and Pacific Ocean, and beneath many continental regions [*Roller and Jackson*, 1966; *Dainty et al.*, 1966]. Similar layers could be seen also in some models of the upper mantle [*Green and Hales*, 1968; *Lewis and Meyer*, 1968]. A

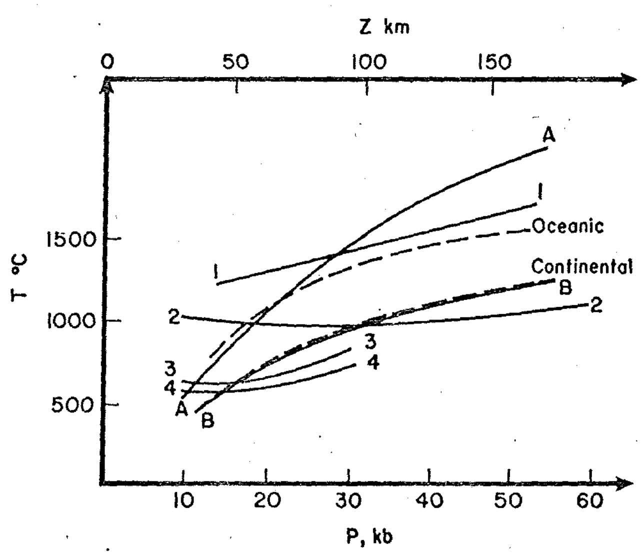

Fig. 4. Temperature in the mantle. *A–A* is the upper boundary; *B–B*, the lower boundary. The broken line represents the Clark and Ringwood geotherms. The numbered lines are: 1–1, dry lherzolite; 2–2, lherzolite + H_2O; 3–3, gabbro + H_2O; and 4–4, granite + H_2O.

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Far from the melting point α is
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...tative explanation of the contradictory data is given below.

Seismic observations give adiabatic velocities of P waves. It was discovered in ionic crystals that the adiabatic value of v_p increases remarkably just before the melting point [*Ubbelohde*, 19[illegible]], owing to rapid growth of thermal expansion α in the course of the approaching melting point. Adiabatic velocity $(v_p)_s$ and isothermal velocity $(v_p)_T$ are related by

$$(v_p)_s^2 = (v_p)_T^2 + \frac{\alpha^2 K_T^2 T}{\rho C_v} \qquad (2)$$

where C_v is the heat capacity at constant volume and K_T is the isothermal bulk modulus. Far from the melting point α is 3×10^{-5}–4×10^{-5}. In the vicinity of the melting point α could be augmented by a factor of 2–3, which gives an additional increase $\Delta(v_p)_s = 0.6$ km/sec near the melting point. The validity of this explanation could be verified by the absence of a high-velocity layer for S waves.

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Convective In...

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A hydrodynamical analysis is given ... viscosity. The solutions are obtained ... layers restricted by rigid walls, and (2) ... fluid. These solutions are then investig... the development of instability are cons... obeys the creep laws. In conclusion, the ... considered.

Convective motions of substances of diff... ...sities constitute the dominant, or one o... ...minant, sources of tectogenesis. A la... ...sity inversion often occurs in the earth... a lighter horizontal layer appears u... ...ated under the heavier one. If the visc... ...rocks is not very large, the convective ... ...y of the Rayleigh-Taylor type [*Ray...*; *Taylor*, 1950] arises. For geotecton... number of properties of the convective ... ...ty can be determined by using a h... ...amical analysis that considers the roc... ...compressible Newtonian fluids of very ... ...cosities.

The Rayleigh-Taylor instability was stu... many authors (*Harrison* [1908], *Char...khar* [1961], and others). Certain prob... were considered for the tectonic applica... for a case when the viscosity of flu... ...th and their motion strongly depends on ...ditions at the boundaries of the layers. In this respect, however, only the formul... the problem was obtained [*Danes*, 196... at last the solutions for the simplest partic... ...ses were found. The most detailed ana... was performed by *Ramberg* [1967]. Ram... ...sidered the instability for two models: ... layers limited by two rigid walls, and ... layers limited by a rigid wall and a ...face. To simplify calculations, the boun... ...ditions were taken to be physically inco... at the interface of the fluids. The horiz... ...ocity components were taken equal to