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Temperature benchmarks in the lithosphere, marking the phase transformations of the matter, and the possibilities of estimating their depths

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Abstract. Extensive research into temperature distribution in the lithosphere and upper mantle revealed it to be highly ambiguous. Of great interest in this context is the distribution of temperature benchmarks in the lithosphere, which mark phase transitions in the individual components of rocks and in the fluids filling their pores. As follows from the percolation theory, the physical properties of the rocks change jumpwise at the discontinuities produced by their phase transformations. This simplifies the interpretation of magnetotelluric and other electrical measurements, enhances the reliability of its results, and facilitates the use of radiolocation, especially in the case of relatively shallow discontinuities. The discontinuities may not coincide in terms of percolation and mechanical conpling, this explaining differences in the results of electrical and seismic measurements.

There is extensive literature on the thermal state of the deep-seated rocks of the Earth, including the papers reporting activities in the frameworks of important international projects. The ideas of temperature variation with depth, evaluated by some geoscientists as highly ambiguous [e.g., *Lyubimova and Feldman,* 1975], are based on the general physical concepts and indirect data. The common approach is to solve a thermal conductivity problem for a model of the crust and upper mantle, which is believed by a particular researcher to be most probable with various assumptions. A discrepancy between the calculations based on different models of the lithosphere averages 100% (of the least). Another approach is to use the variation of the electrical conductivity of rocks with depth, provided by the MTS, MVS, and other methods of electromagnetic sounding with the controlled sources of an electromagnetic field and compare it with the results of the experimental measurements of the variation of electrical conductivity with temperature and pressure. The most comprehensively developed models of a normal (standard) resistivity section, based on MTS data [*Van'yan and Shilovskii,* 1983; *Zhamaletdinov,* 1990], suggest a gradual growth of the electrical conductivity of rocks with depth with the mere variations of its gradient at the internal boundaries of the section. In this case, too, the order of conductivity uncertainty is great enough to produce differences between the models of different authors, amounting to one or two orders of magnitude. Another potential feasibility of estimating the temperature distribution in the Earth's crust is to use geophysical methods for locating temperature benchmarks – the boundaries at which a phase transition, accompanied by a change of some physical property of rocks, takes place, as some of the rock components or the interstitial solution (generally fluid) attains a phase-change temperature. Examples of these temperature benchmarks are the transformation of water to ice or to some supercritical state, the partial melting of rocks (basalt) in the asthenosphere, the transformation of ferromagnetic substances to paramagnetics as the temperature attains the Curie point, the transition of olivine to spinel, etc. The difficulty is that, for example, the temperatures of water transition to ice, or to a supercritical state [*Smith,* 1968], depend on the composition and concentration of substances, dissolved in it, and on pressure. Uncertainty of the same kind is encountered, for a number of reasons, in the cases of a partial melting zone in the asthenosphere and in the zone of olivine transition to spinel. In accordance with the hypothesis of the pyrolite composition of the Earth's mantle, and with the assumption of $Fe/(Fe + Mg) \approx 0.1$, the temperature of this phase transition under pressure of 135 kb must be $1600° \pm 50°C$

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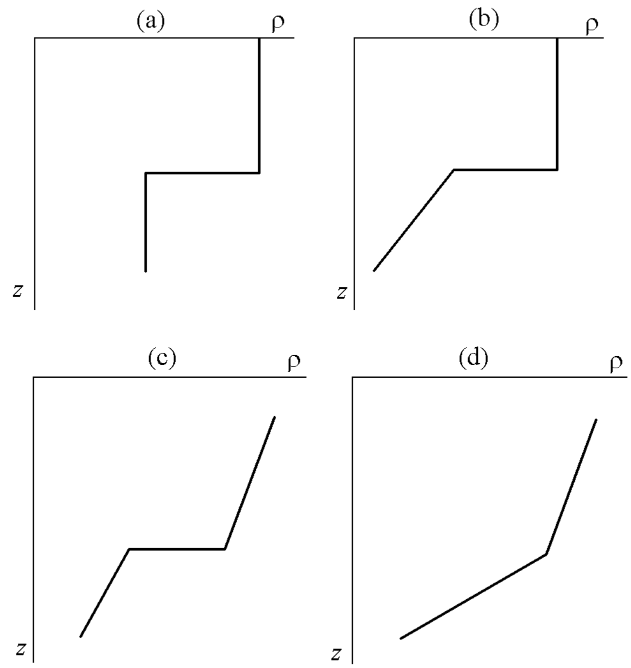

Figure 1. See the text for the explanation.

[*Zharkov,* 1978]. This temperature will be different in the case of any deviations of the olivine composition from the composition mentioned above. *Akimoto and Fujisava* [1965] found the temperature of the phase transition of olivine to spinel at 43.8 kb to be ~1170°C. It follows that in the general case we are dealing with the temperature estimation of a particular phase transition from above or from below. Another important condition is the character of a change in the studied property of the rock during its phase transition: jumpwise, as follows from the general phase transformation theory, or the variation gradient of a given property changes with depth, as it is assumed for the upper boundary of the asthenosphere [*Van'yan and Shilovskii,* 1983].

The most widely used temperature benchmark is the zero isotherm based on the thickness of frozen rocks [*Kalinin and Yakupov,* 1989; *Yakupov,* 2000] and the isotherm of the Curie point of magnetite based on the depth of the bottoms of the magnetized bodies [*Bulina,* 1970; *Volk et al.,* 1977a, 1977b]. A contact between thawed and frozen rocks is generally marked by the abrupt changes of electrical conductivity, dielectric constant, and, in the case of coarse-grained loose deposits and coarse-clastic hard rocks, of seismic velocity, or, in the general case, of electrical and acoustic impedances. The regions with fresh and salt subpermafrost water can be distinguished by the absence or presence of a stochastic correlation between the thickness of the permafrost rocks and the conductivity of a subpermafrost layer [*Kalinin and Yakupov,* 1989]. Therefore, in the regions of the coexistence of frozen rocks and fresh subpermafrost water, the position of the first of the above mentioned temperature benchmarks – $t = 0°$C – can be located reliably by the methods of electrical prospecting under favorable conditions with a mean relative error of ~10%. We succeeded to solve this problem for one of the largest water-bearing structures with salt subpermafrost water, the Olenek artesian water basin. We derived a temperature regression equation for the lower contact of the frozen rocks as a function of their thickness [*Kalinin and Yakupov,* 1989].

Another difficulty is the fact that the character of the variation of a particular physical property of the rocks during their phase transformation is not clear at first glance. One can assume the following variants for the behavior of electrical resistivity with depth in the vicinity of or at the boundary produced by a phase transition in any constituent of the rock (Figure 1). The simplest case (Figure 1a) is, for example, a change of resistivity at the lower boundary of a lithologically homogeneous frozen rock sequence with subpermafrost fresh water, that is, at a zero temperature [*Yakupov,* 2000], or a change of the magnetic permeability of ferromagnetics at a Curie temperature. The electrical conductivity of rocks increases at depth with a temperature growth. It is believed that with the formation of a partial melting zone in the asthenosphere, its conductivity grows with depth more rapidly, so that its upper surface is marked by an abrupt (jumpwise) growth of a conductivity gradient [*Van'yan and Shilovskii,* 1983], this being the most complex case (Figure 1d). In view of its generality, we will consider this case using the theory of phase transitions and the idea of a percolation theory. A zone of partial melting arises in the top of the asthenosphere because of basalt melting. The melt portion (2 to 30% of the volume after *Ringwood* [1975]) is believed to grow with temperature, that is, with depth. It is known, however, that with all other factors being equal, any phase transition of a particular material from one state to another takes place at a constant temperature. In the case of a high heat flow, that is, in the case of fairly rapid heating or cooling, a transition zone can arise, where both a solid and a liquid phase coexist. Excluding the cases of the origin of magma chambers in the regions of volcanic activity, this phenomenon does not take place in the deep lithosphere, and the thermal state of the latter can be taken to be quasistationary. Therefore a partial melting zone in the asthenosphere is bound to have a distinct interface, where some of its physical properties change jumpwise. This interface shows pressure and temperature variations, and hence changes in resistivity, but not in the melt content, at least in the case of a two-component material.

It is of interest to consider electrical resistivity at the boundary of a partial melting zone and in both of its sides in terms of a percolation theory. The rocks of the lower lithosphere and asthenosphere are polycrystalline formations. The arrangement of individual components in them is the compact packing of their chaotically distributed particles of different size and form. In this case the electrical conductivity of such a fairly large 3-D system can be described by the formula

$$\gamma = \gamma_2 (x - x_c)^t + \gamma_1 (1 - (x - x_c)), \quad x \leq x_c \ , \qquad (1)$$

where γ_1 is the conductivity of the asthenosphere above the partial melting zone, γ_2 is the conductivity of the melt, x is the relative melt content from 0 to 1, x_c is the melt critical value, at which melt percolation takes place, this value being 0.25 in our case, and t is the critical index of electrical conductivity. Formula (1) is valid, as a first approximation, with an accuracy of a factor for γ_1 and γ_2 (we are interested in the behavior of the electrical conductivity of rocks in the vicinity of or at the melt percolation). Taking into account the sinuosity of the framework of an infinite melt cluster, proved for 3-D media, $t = 1 + \nu$, where ν is the index of the correlation radius, equal to 0.8–0.9. It should be added that the formation of an infinite cluster is marked by γ tending to be equal to γ_1. Hence, we arrive at

$$\gamma = \gamma_2(x - 0.25)^{1.85} + \gamma_1(1 - (x - 0.25)), \quad x \leq 0.25 \ . \quad (2)$$

Figure 2 shows a logarithmic curve of the variation of the systems' electric conductivity as a function of temperature, where the dots show the $\log \gamma = f(x)$ values in accordance with formula (2). It should be emphasized that electric conductivity varies further with depth, following pressure and temperature variations: the melt portion in the resulting two-phase system remains constant and can change (jumpwise!) only when another component of the asthenospheric material begins to melt. To sum up, the interfaces produced by the phase changes of the material, when the percolation threshold is achieved, are marked by the jumpwise changes of at least some of the physical properties of the material, and by the $\gamma = \varphi(z)$ function having the form displayed in Figure 1c, in the case of electric conductivity. This distribution of the electric conductivity of lithospheric rocks with depth simplifies the interpretation of magnetotelluric and magnetic variation measurements and also of the measurements made using the controlled sources of the electromagnetic field. Moreover, this facilitates the use of radiolocation where the appropriate equipment is available. The main trends of research here are the use of high-power MHD generators [*Velikhov and Volkov,* 1982] to generate pulsed electromagnetic fields; the use of the strength of the magnetic field as the parameter of interest, and the use of high-precision magnetometers based on the Josephson effect as receivers to record the arrival times of reflected waves. This will enhance highly the accuracy of determining the depths of the discontinuities in question.

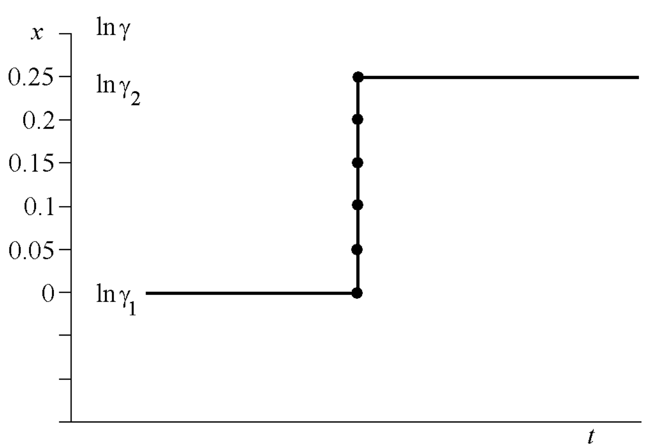

Figure 2. Electrical conductivity of the asthenosphere for the case of a one-phase transition. The Y-axis also shows a scale for the relative content of melt in the rock from the zero to the threshold value of x_c.

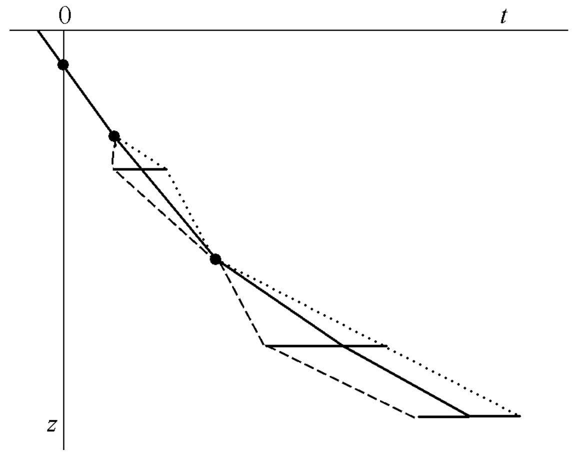

Figure 3. Schematic distribution of temperature in the lithosphere from its values at the interfaces produced by the phase transitions of the material: 1 – ice–water transition, 2 – Curie point for pyrrhotite, 3 – transition of the rock to an above-critical state; 4 – Curie point for magnetite; 5 – transition of olivine to spinel; 6 – the upper boundary of a partial melting zone in the asthenosphere.

The main contribution to error in plotting the temperature t variation with depth $t = f(z)$ will be made by the determination of the temperature of a particular phase transition, which depends on the contents of other rock components and pressure. This temperature is known for the freezing of fresh water in the pores of the rocks and for the second-order phase transitions accompanied by the transformation of ferromagnetic minerals to paramagnetic ones, magnetite and pyrrhotite. Other cases will deal with the estimation of the phase-transition temperature from below or from above depending on the contents of volatiles and other components in the rocks. All of these estimates will be refined with time. The hypothetic function $t = f(z)$ will have the form similar to that shown in Figure 3, on the assumption that all interfaces produced by the phase transitions of the material were located in one region or under similar conditions. The dotted lines in Figure 3 show the positions of the interfaces with the known phase transition temperatures (water – ice; pyrrhotite, magnetite paramagnetics). The horizontal line segments in Figure 3 are conventionally equal to the range of the potential temperature variation for a particular phase transition: of water to its above-critical state, of olivine to spinel, of basalt melting; the dashed lines depict the $t = f(z)$ estimates from below, the dotted lines, from above. In the general case, after the formation of melt from one of the rock components, its segregation and changes in the contents of or in the ratio between some other rock components, e.g. water, or in pressure, may involve the melt-

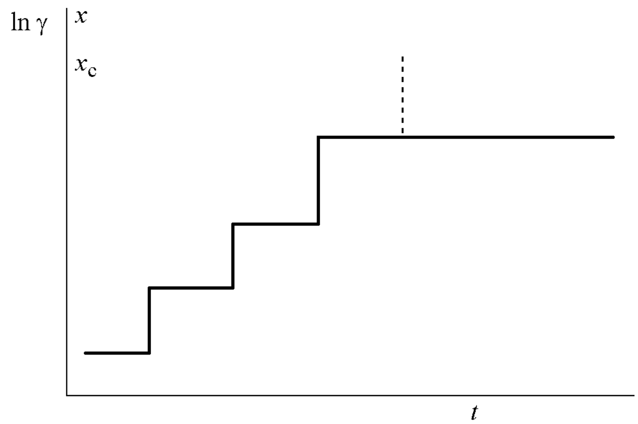

Figure 4. The electrical conductivity of the asthenosphere as a function of temperature for the case of material phase transitions in it with the attainment of the x_c critical value and melt percolation during one of the subsequent transitions.

ing of another rock component, then of another, and so on. Each time during a next phase transition in the rock components (meaning their melting), the total volume of the melt will increase jumpwise, while the temperature of the rock as a whole will remain constant till the end of the next phase transition. Melt percolation may take place at one of these phase transitions, and not necessarily at the first one, as depicted conventionally in Figure 4. The lower boundary of the partial melt zone will obviously be controlled by the depletion of the relatively low-melting components of the asthenospheric substance.

The threshold of percolation, as a constant of transportation processes, and the threshold of mechanical coherence in the system arising as a result of a phase transition may not coincide, and in this case the positions of the interfaces determined by electrometric and seismometric methods will be different (see, for example, [*Yakupov,* 2000]). Where the content of a new phase is small, the mechanical coherence of the initial rock remains intact, as does its homogeneity in terms of the related parameters.

Conclusion

1. The positions of the interfaces produced in the lithosphere by phase transitions in the individual components of the rocks, or in their interstitial fluids, control the positions of its temperature benchmarks. The temperatures of some phase transitions are known. In the case of the others it is controlled by the ratio or contents of the other rock components and pressure.

2. Some of the physical properties change jumpwise at the interfaces produced by the above mentioned phase transitions. Three models are proposed for their distribution in depth: models a, b, and c in Figure 1. These models can be used to determine the positions of these interfaces, in the case of their sufficient contrasts, using magnetotelluric sounding, magnetic variation measurements, and soundings using controlled field sources. This simplifies the interpretation of the results and makes them more reliable. Theoretically, this problem can be solved using radiolocation. Some technical means, though limited in their performance, are available.

3. Where the content of a new phase is sufficient, the interfaces may coincide in terms of percolation and mechanical coherence, and in this case its position can be located by electrometric and seismic methods. Where these thresholds differ, but have been achieved, the positions of the interfaces indicated by electrometry and seismic methods are different: the interface determined by electrometry must lie higher. Where the threshold of mechanical coherence is not attained, the physical properties of rocks remain intact.

4. The materialization of the ideas advanced in this paper will hopefully enhance our knowledge of the temperature distribution in the lithosphere, for instance, in the form presented in Figure 3.

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