TEMPERATURE GRADIENTS AT THE CORE-MANTLE INTERFACE*

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Abstract. Heat flowing out of the core must flow into the mantle. If the Earth's magnetic field is owing to adiabatic magnetohydrodynamic circulation of the outer core, whole mantle convection or melting at the core mantle boundary is required to keep the inner core from becoming isothermal, thereby preventing adiabatic circulation.

Alternatively, the outer core fluid must have some unexpected and exotic property such as an extremely low coefficient of thermal expansion and resultant low Gruneisen's parameter.

1. Introduction

Stacey (1972a) recently suggested that if the liquid in the outer core of the Earth is indeed adiabatic, heat flowing down the adiabat would have to escape into the mantle, and that unless special heat sources are postulated for the core and special heat sinks are postulated for the mantle, the core of the Earth would soon become isothermal. We have come to conclusions not too different from those of Stacey, and hope in this paper to reinforce his arguments with some different numerical calculations.

Rates of heat generation and heat escape from the core of the Earth into the mantle have been previously examined by Bullard (1950), Bullard and Gellman (1954), Verhoogen (1961) and Stacey (1969). These authors selected a core model, estimated the temperature gradient in the outer core, the conductivity of the outer core, and the core temperature. From these estimates they then calculated the heat flux into the mantle. The parameters used by these authors, for various models, are shown in Table I. As can be seen, the estimates of core temperature, thermal conductivity, and the core gradient range widely.

Bullard (1950) considered this topic at the earliest date. He estimated the thermal conductivity of liquid iron to be 0.19 cal cm^{-1} s^{-1} °C by averaging his and Elsasser's estimate of the electrical conductivity of the outer core. Bullard notes that even with a deep mantle gradient of 3 °C km^{-1}, which is thirty times our current estimate of the deep mantle gradient, the calculated heat flux from the core is fourfold greater than can escape by conduction. He points out that this problem is solved by whole mantle convection but notes that if his estimated conductivities and gradient hold, approximately 50% of the known surface heat flow is heat escaping from the core.

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TABLE I

Heat flux from the core

	Core gradient ° km^{-1}	Conductivity outer core Cal cm^{-1} s^{-1} deg^{-1}	Flux Cal cm^{-2} s^{-1}	Core temp. °C
Bullard (1950)	1.1	0.19	2×10^{-6}	10000
Bullard and Gellman (1954)	0.26	0.10	2.6×10^{-7}	5000
Verhoogen (1961)	0.22	–	2×10^{-7}	2300
	0.51	0.10	5×10^{-7}	3000
Stacey (1969, 1972a, b)	0.2	0.01	2×10^{-8}	3700
	0.9	0.067	6×10^{-7}	3100
Kennedy and Higgins	0.29 (m. p. Gradient)	0.20	6×10^{-7}	3700
(this paper, 1972)	1.2 (adiabatic gradient)		2.4×10^{-6}	3700

This problem arises from his assumed very high core temperature and, it appears to us, that a similar problem would arise from any model assuming very high core temperatures, such as that of Verhoogen (1972).

Bullard and Gellman (1954) later compute the heat flux from a core of sharply lower temperature and assume a lower thermal conductivity for the outer core. The computed heat flux drops by one order of magnitude and the heat flux problems associated with the assumed high core temperature are sharply diminished.

Verhoogen (1961) discussed the heat balance of the Earth's core. Using the Bullard and Gellman (1954) value for thermal conductivity of iron, 0.1 cal cm^{-1} deg^{-1} s^{-1}, he computes that a mantle gradient of 1.45 deg km^{-1} is required to dissipate the heat from a core at 2300 °C and a gradient of 1.9 deg km^{-1} is required for a core at 3000 °C. Inasmuch as it is 2900 km to the core mantle boundary, such gradients are inconsistent with the assumed low core temperatures, unless, of course, whole mantle convection or melting at the core mantle boundary is assumed.

Stacey (1969) presented a core model in which he computed an extremely low heat flux into the mantle. The major feature of this model, however, was an assumed thermal conductivity for liquid iron of 0.01 cal cm^{-1} deg^{-1} s^{-1}. This value is less than half of the conductivity we now estimate for silicates of the deep mantle and is about $\frac{1}{20}$th of the value computed from shock measurements of the resistivity of iron at high pressures (Mitchell and Keeler, 1971). Stacey (1972a, b) reconsidered the heat flux from the core, using a higher adiabatic gradient and an increased thermal conductivity for iron (Gardiner and Stacey 1971), with sharply different conclusions.

2. Discussion

Let us first examine the temperature gradients that would be expected in the outer core of the Earth provided there were no heat sinks in the mantle other than the heat that can escape by flowing down the temperature gradient in the deep mantle. Under these circumstances, assuming no circulation, the steady state ratios of the temperature gradients of the liquid outer core of the Earth and the deep mantle will be inversely proportional to their thermal conductivities. Thus we need to estimate the temperature gradient in the deep mantle, the thermal conductivity of the deep mantle, and the thermal conductivity of the liquid outer core of the Earth.

A large number of Earth models showing mantle temperature gradients based on a variety of assumptions have been published over the last few years. Perhaps the most complete group of models is that published by MacDonald (1959). These models have been calculated using various assumptions as to the distribution of radioactivity and opacity. Most of them show the deep mantle to be essentially isothermal, i.e., deep mantle convection could not take place and no heat can be escaping from the core of the Earth. However, the model showing the steepest deep mantle temperature gradient is the one in which MacDonald assumes that radioactivity is uniformly dispersed through the Earth and the deep mantle rocks are of high opacity. In this model, the temperature gradient near the mantle-core boundary amounts to approximately 0.4° km^{-1}. These temperatures closely approach the extrapolated melting curve for diopside. MacDonald concludes that with higher gradients than this, the Earth would be largely molten. The limit on the maximum permissible temperature gradient is thus set by the beginning of melting of mantle rocks.

Kennedy and Higgins (1972a) recently attempted to reestimate the liquidus, the solidus, and the adiabatic gradient in the Earth's mantle. Their results are shown in Figure 1. Their estimate takes into account the fact that eutectic troughs deepen with pressure. The deep mantle is certainly a multi-component system with melting temperatures set by the eutectic minima. They estimate the solidus and maximum deep mantle temperature gradient to be 0.10–0.15 deg km^{-1}. In additon, they have computed the adiabatic gradient of the mantle, assuming the temperature of the deep mantle is that of the solidus. A striking feature of these results is that the adiabatic and the solidus curves essentially are superimposed over the lower third of the mantle. This feature of their results suggests that convection is essentially restricted to the upper two-thirds of the mantle, but it does not absolutely preclude whole mantle convection as the uncertainties in the calculations are too great.

We next need an estimate of the thermal conductivity of rocks of the deep mantle. The task of estimating the thermal conductivity of an assemblance of minerals, either silicates or oxides, in the deep mantle of the Earth is an exceedingly difficult one. However, it has been recently investigated at great length from both an experimental and theoretical point of view by Schatz (1971). Schatz concludes, "Results for single crystal and polycrystalline forsterite-rich olivines indicate that, even in relatively pure, large crystals, radiative conductivity does not increase rapidly with temperature. The predicted total conductivity at 500 kms depth in the Earth's mantle is less than twice the surface olivine value of about 0.012 cal $cm^{-1} s^{-1} deg^{-1}$(0.05 J $cm^{-1} s^{-1} deg^{-1}$)". We thus estimate a value of 0.025 cal $cm^{-1} s^{-1} deg^{-1}$ as the value for thermal conductivity in the deep mantle. Hopefully, this is not in error by more than a factor of 2.

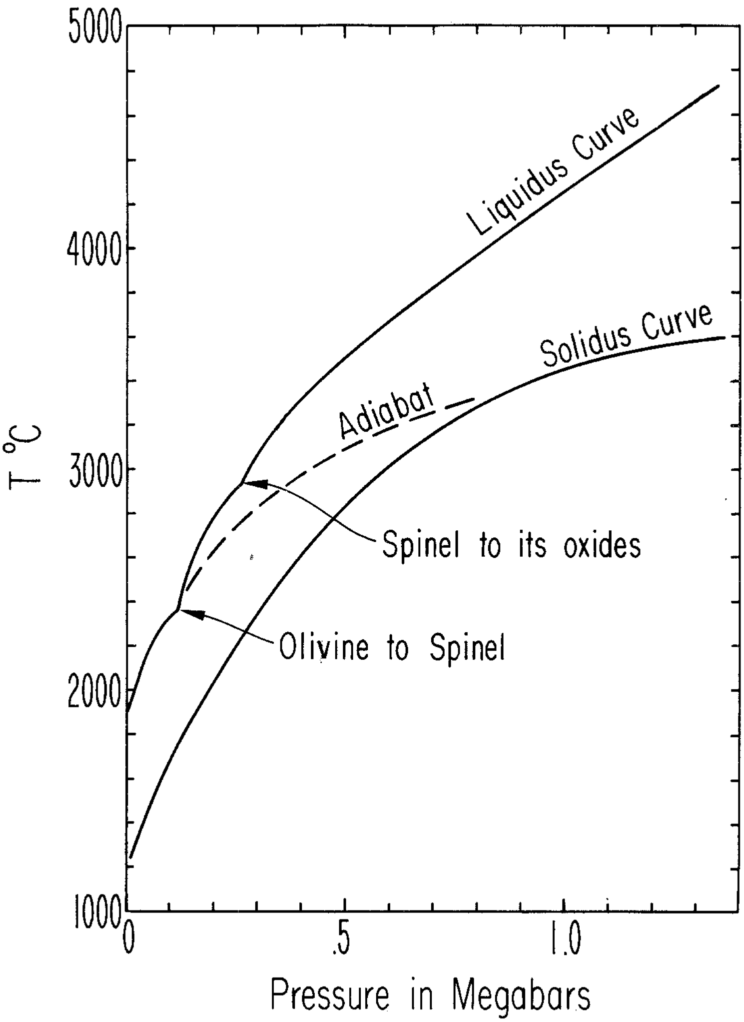

Fig. 1. Estimated solidus, liquidus and adiabate for the mantle of the Earth.

We need now to make some estimates of the thermal conductivity of liquid iron under conditions similar to those of the outer core, i.e., temperatures circa 4000 K and pressures of 1.4–3.1 mbar. Measurements of the electrical conductivity of iron have been made by Mitchell and Keeler (1971) in shock experiments.* The results show that above 200 kbar the product of electrical conductivity and temperature of solid iron increases linearly with pressure up to 1.4 megabar, the limit of the experimental data. The electrical resistivity of most metals increases by a factor of approximately 2 on melting. Assuming the Wiedemann-Franz law that thermal conductivity is proportional to the product of electrical conductivity and temperature, we derive that

* These published results have an inadvertent factor of 10 error in the conductivity scale.

thermal conductivity of iron at core pressures is 0.14 cal cm^{-1} s^{-1} deg^{-1} P (Mb). Pressure ranges from 1.4 to 3.1 megabar in the outer core. Thus thermal conductivity should range from 0.2 to 0.4 cal cm^{-1} s^{-1} deg. This number is essentially the same as that estimated by Bullard (1950). These very rough estimates suggest that the thermal conductivity of liquid iron in the outer core is at least 10fold that of the silicates near the core-mantle boundary. Even though there are substantial uncertainties in these estimates, we can see no escape from the conclusion that thermal conductivity in the liquid outer core of the Earth is very much greater than in the deep mantle.

Assuming there are no heat sinks other than thermal conductivity we must conclude from the foregoing that the gradient in the outer core is one tenth that of the deep mantle or about 0.01–0.015 deg km^{-1}. According to this model the core is essentially isothermal and the above rather simple-minded arithmetic suggests a difference of circa 25 °C–35 °C through the outer core. However, this result is in sharp conflict with several other analyses of core temperature distribution and would completely inhibit any kind of adiabatic radial mixing of the core.

In a previous paper Higgins and Kennedy (1971) suggest that a temperature difference of circa 1250 °C is required for the outer core to circulate adiabatically. They also suggest a melting point difference of circa 500 °C for an essentially pure iron core. They came to the conclusion that temperatures in the outer core were probably distributed along a melting point curve of a multi-component iron rich system. Due to lowering of the melting points in the multi-component system, a core gradient somewhat less than that of pure iron was suggested. The calculation of the adiabatic gradient is consistent with the assumption of a 'normal' Gruneisen gamma for liquid iron of circa 1.66 and the observed seismic velocities in the outer core. If the core were as nearly isothermal as heat balance requires, a Gruneisen gamma of circa 0, would be required for the outer core to circulate adiabatically. This implies that at core temperatures and pressures the coefficient of thermal expansion for the core fluid would be essentially zero. It is also apparent that if Gruneisen's gamma were near that of any of the recent estimates for the liquid in the outer core, the outer core must be stratified thermally in a very stable way, and this would impose a powerful inhibition on magnetohydrodynamic circulation in the outer core of the kind implicit in recent discussion of the origin of the Earth's magnetic field.

The way out of this conclusion that the core of the Earth is essentially isothermal is by postulating a heat source in the core of the Earth and a heat sink at the core-mantle boundary. Following Verhoogen (1961) and others, we can propose that the freezing of iron at the inner core–outer core boundary might be a sufficient heat source, and suggest that KS dissolved in the core fluid, following the suggestions of RamaMurthy and Hall (1972) might be an additional heat source. Two different heat sinks in the deep mantle may be proposed. We can assume that convective circulation of deep mantle rocks carries sufficient heat away so that a high temperature gradient can be maintained in the core; or we can propose that melting at the core-mantle boundary and upward escape of liquid along a melting point gradient serves as a heat sink.

Let us examine a model. Assume, for the moment, that the temperature of the inner core–outer core boundary was 3700 K and that the temperature of beginning of melting of silicate in the deep mantle was 3500 K. Let us also assume that temperatures in the liquid outer core were initially distributed along a freezing point gradient of 300–500 °C through the liquid outer core. In the absence of whole mantle convection, heat would flow down this gradient and the temperature at the core-mantle boundary would rise until the melting temperature of the silicate at the postulated 3500 °C was reached. This temperature difference of 300 °C could then be maintained.

Assuming thermal conductivity in the outer core of 0.2 cal cm^{-1} s^{-1} deg^{-1} and a melting point gradient of circa 0.1 deg km^{-1} in the outer core, we compute that 2×10^{-7} cal cm^{-2} s^{-1} will be flowing by thermal conductivity down the melting curve and into the mantle. This rather great loss of heat can just be sustained by freezing at the inner core-outer core boundary. We estimate that if the inner core has been freezing at a uniform rate over the last 4.5 billion years, and assuming a uniform drop in temperature of circa 100 °C, during the same time the rate of growth will sustain a flow of circa 2×10^{-7} cal cm^{-2} s^{-1}. Thus no heat deficit appears and we do not need to involve a radioactive heat source to maintain a core gradient.

If one postulates a larger temperature gradient through the outer core, a source of heat other than that supplied by freezing, such as KS in the core, is required. It seems most unlikely, however, that an outer core fluid, with a temperature difference between the inner core–outer core boundary and the core-mantle boundary of only 200 °C could be adiabatic. This, again, would imply a Gruneisen parameter for the liquid outer core material of circa 0.25 which seems improbably low (see Figure 2). Melting at the core mantle boundary with upward escape of liquid would serve as a heat sink.

Stacey and others have proposed whole mantle convection for the transportation of this heat to the surface. However, reference to our Figure 1 shows that the adiabat in the mantle is essentially superimposed on the solidus over the lower one-third of the mantle. Thus, to the extent that our curves are right, whole mantle convection appears unlikely. We note, however, that the solidus curve that we have estimated is not too different in temperature at the mantle-core boundary from the estimated temperature of melting of iron. We estimate the solidus of the deep mantle rocks to be circa 3500 °C–3600 °C at the core-mantle boundary; the estimated temperature for the melting of the iron mixture at the same depth is only slightly larger. Thus, it seems that the requisite heat sink can be formed by the melting of silicate or oxide materials near the core-mantle boundary and upward migration along the melting curve of liquid. This implies, of course, that the temperature gradient through the mantle is that of the solidus curve. There is some seismological evidence which tends to support the possibility of melting at the core-mantle boundary. Jordan and Anderson (1972) and Phinney and Alexander (1966) report a substantial decrease in shear velocity immediately above the core-mantle boundary. The low velocity zone in the upper mantle is commonly attributed to partial melting and a similar suggestion might offer an explanation for the seismic anomalies at the core-mantle boundary.

It appears that temperatures in the outer core, distributed along the melting point

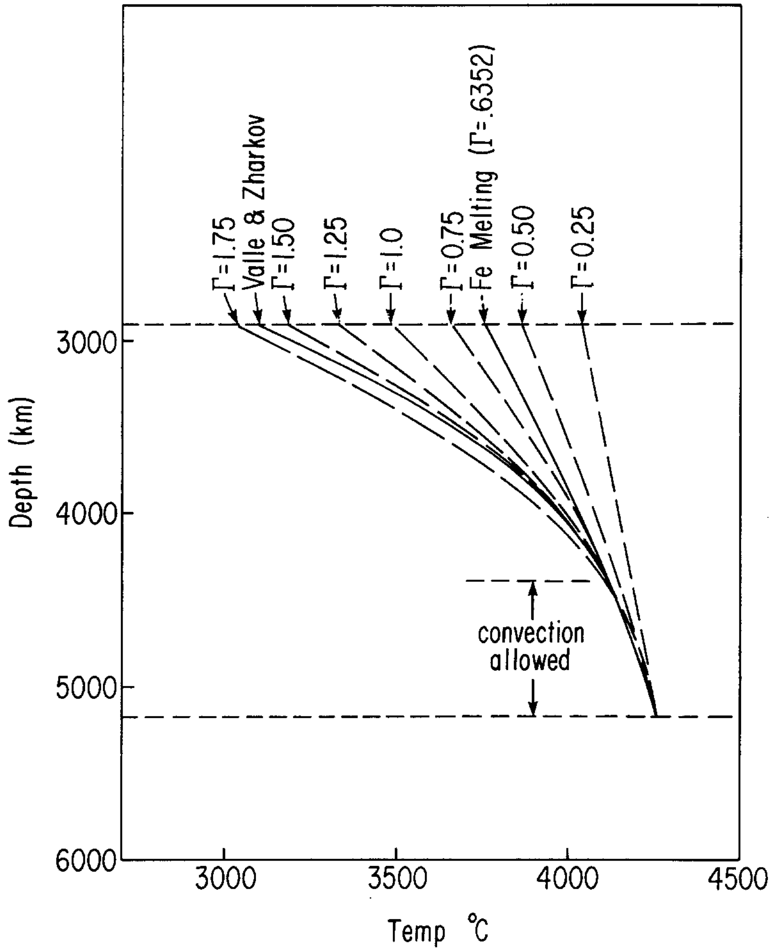

Fig. 2. A melting curve for iron and calculated adiabats using various assumptions as to Gruneisen's parameter.

gradient of 0.1° km^{-1}, pose no insuperable problems but do require melting at the core mantle boundary or whole mantle convection. Inspection of Table I, however, shows that it is less probable for the temperatures to be distributed along our computed adiabatic gradient of 1.2 deg km^{-1} at the core mantle boundary based on a 'normal' Gruneisen gamma of 1.66 for iron. Flow of heat down this gradient would be 2.4×10^{-6} cal cm^{-2} s^{-1} which is about half the total heat flow observed at the surface of the Earth and would require the partitioning into the core of several hundred parts per million of potassium.

3. Conclusions

Two conclusions appear. If the Earth's magnetic field is owing to adiabatic magneto-hydrodynamic circulation we must require that the core fluid have an unexpectedly low value of thermal expansion with a resulting very low Gruneisen's gamma. Secondly, we require whole mantle convection or melting at the core mantle boundary and upward escape of liquid in order to provide a heat sink so that the appropriate gradients may be maintained across the core. If there is no convection or melting in the deep mantle then the temperature gradient across the core must be 25–35 °C.

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