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# Heat Production and Thermal Conductivity in Mainland Portugal

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## ABSTRACT

Temperature distribution within the Earth affects a great deal of phenomena and properties of earth materials. To understand the geophysical and geodynamic behaviour and evolution of the interior of the earth is essential knowing its temperature distribution and thermal regime. One of the most important quantities to estimate the temperature distribution inside the earth is heat flow density. In principle, with heat flow density values measured at the surface of the earth and integrating the heat conduction equation, with the appropriate boundary conditions, it is possible to estimate the temperature distribution down to the Moho and deeper. However, temperature distribution in the lithosphere is highly dependent on the thermal conductivity and heat production of the geologic formations. In this contribution, a compilation of laboratory measurements of thermal conductivity and heat production for some geological formations that outcrop in Mainland Portugal are presented and are used to extrapolate measured surface temperatures to the crust and, in some cases, to the lithosphere.

Heat flow density in Mainland Portugal varies from 42 to 115 mW/m$^2$. These values are similar to heat flow densities obtained in other European Hercynian regions and allow inferring that temperatures at Moho depth should not be higher than about 700 ºC. One and two dimensional geothermal modelling suggests that in southern Portugal heat production in the crust must be as high as 4.8 µW/m$^3$, which is higher than previously suspected for the area and higher than measured values for surface rocks.

## 1. INTRODUCTION

The temperature distribution inside the earth is of great importance for understanding geophysical and geological phenomena; it also influences most of the values of the physical properties of the rocks. Electrical, magnetic and seismic properties of earth materials as well as density, and chemical and rheological processes are highly dependent on temperature. Therefore it is important to know the temperature distribution and the thermal regime inside the earth. At the earth's surface temperature can be measured directly in boreholes; for deeper regions temperature has to be extrapolated using heat flow density estimated at the surface. However, for depths deeper than about 10 km, extrapolation of temperature is difficult as a result of an increased uncertainty in the knowledge of thermal conductivity and heat production of the geological formations. For estimating temperature for crustal depths, besides knowing the heat flow density at the earth's surface, it is imperative to estimate the thermal conductivity and the heat production for deep seated rocks from values measured at the surface or near the surface. For the lithosphere the estimation of the heat flow density is relatively straightforward because heat transfer is essentially by conduction; for the asthenosphere the thermal regime is more complicate due to the fact that heat transfer is also by convection. In this paper we will only be discussing the thermal regime in the crust and lithosphere.

The basic quantity to study the thermal regimes in the crust and lithospehere is the heat flow density defined by eq. (1)

$$Q = k \cdot grad\, T \tag{1}$$

where $Q$, $k$, and $gradT$ are the heat flow density, the thermal conductivity, and the geothermal gradient, respectively.

In the general case, heat flow density and geothermal gradients are vectors and the thermal conductivity is a tensor. Eq. (1) allows to calculate the heat flow density but does not give any information about the temperature distribution within the earth. That information can though be obtained, using the thermal conduction equation that, in the most general situation, can be written as:

$$-div\,(-k \cdot grad\, T) + A = \rho \cdot c \cdot \frac{\partial T}{\partial t} \tag{2}$$

which allows for time changes of temperature distribution and where $div$ is the differential operator divergence, $A$, $\rho$, $c$, and $t$ are the heat production per unit volume, the density, the specific heat, and the time, respectively.

When temperature changes in the vertical direction only and the lithosphere can be assumed as homogeneous and isotropic in terms of thermal conductivity and heat production, eq. (2) can be simplified to:

$$\frac{\partial^2 T}{\partial z^2} + \frac{A}{k} = \frac{1}{\alpha} \cdot \frac{\partial T}{\partial t} \tag{3}$$

where $\alpha$ is the thermal diffusivity which represents the facility with which heat diffuses in a given material.

In geothermal studies it is also assumed that thermal processes are very long in time (stationary thermal regime) and eq. (3) can be simplified again to be written as Poisson partial differential equation:

$$\frac{\partial^2 T}{\partial z^2} = -\frac{A}{k} \tag{4}$$

By appropriate choice of boundary conditions, it is possible to solve eq. (4) to estimate the vertical temperature distribution and so estimate the thermal regime in the crust and/or lithosphere. Solving eq. (4) is, however, not simple; as a matter of fact, rock thermal conductivities and heat production by radioactive decay are parameters difficult to estimate. Thermal conductivity varies with temperature and pressure, and heat production varies with the distribution of thorium, uranium, and potassium which is highly variable in crustal rocks. Solutions of eq. (4) can only be obtained if the distribution of thermal conductivity and heat production is known or estimated for the crust or lithosphere which, in many cases is not.

For shallow geothermal studies it is usual to assume that thermal conductivities do not change with depth. Furthermore, it is also assumed that heat production is described by one of the following models: the step model of Roy et al. (1968) and the exponential model of Lachenbruch (1968). In the first model heat production is assumed to be constant to a depth D; in the second the heat production A(z) is assumed to change with depth according to the equation:

$$A(z) = A_0 \cdot e^{-\frac{z}{D}} \tag{5}$$

where $A_o$ is the heat production at the earth's surface, $z$ is depth and $D$ has the dimension of a distance (depth) and characterizes the vertical distribution of the heat sources.

Using the step model for heat production, the solution of eq. (4), with the appropriate boundary conditions, is:

$$T(z) = T_s + \frac{Q_s}{k} \cdot z - \frac{A_0}{2k} \cdot z^2 \tag{6}$$

Using the exponential model with the appropriate boundary conditions, the solution of eq. (4) is:

$$T(z) = T_s + \frac{q}{k} \cdot z + \frac{A_0 D^2}{k} \cdot \left( 1 - e^{-\frac{z}{D}} \right) \tag{7}$$

where $Q_s$ and $T_s$ are the heat flow density and the temperature, respectively, at the earth's surface, and $q=Q_s - A_o D$.

Eq. (6) can be used to calculate the temperature distribution as a function of depth for a layer of thickness $\Delta z$, thermal conductivity $k$, heat production $A_o$, and surface heat flow density $Q_s$. In case there are several layers of thickness $\Delta z$, with different values for those quantities, the temperature $T_B$ and the heat flow density $Q_B$ at the base of each layer can be expressed by the following equations:

$$T_B = T_T + \frac{Q_T}{k} \cdot \Delta z - \frac{A_0}{2k} \cdot (\Delta z)^2 \tag{8}$$

$$Q_B = Q_T \cdot A_0 \cdot \Delta z \tag{9}$$

where $T_T$, $Q_T$ , and $Q_B$ are the temperature, the heat flow density at the top and at the bottom of the layer, respectively, and k and $A_o$ are the thermal conductivity and the heat production in the layer, respectively.

For a pile of layers the calculation of the temperature distribution as a function of depth is made in an iterative way, layer by layer. Eqs. (6) to (9) have been used by Correia and Jones (1997) and Correia and Ramalho (1999) to construct one-dimensional geothermal models for southern Portugal; eq. (2) has been used to construct two-dimensional models. Figure 1 shows a location map for boreholes used in this study, while Figure 2 presents a two-dimensional geothermal model for southern Portugal.

## 2. THERMAL CONDUCTIVITY AND HEAT PRODUCTION IN MAINLAND PORTUGAL

The temperature distribution inside the earth is mainly controlled by the thermal conductivity and the heat production of the geological materials. For non-stationary geothermal studies thermal diffusivity is also an important property to take into consideration; however, it will not be considered here.

### 2.1 Thermal Conductivity

Thermal conductivity in the lithosphere is basically controlled by the chemical composition of the rocks and, to a lesser degree, by temperature and pressure. Even though it is usual to consider that thermal conductivity is constant in the crust, experimental data indicates that temperature and pressure should be taken into account when calculating the temperature distribution in the earth. The variation of thermal conductivity with depth and temperature is given by the following equation:

$$k(T, z) = k_0 \cdot (1 + c \cdot z) / (1 + b \cdot \Delta T) \tag{10}$$

where $T$ is the temperature in °Centigrade, $k_o$ is the thermal conductivity measured at 0 degrees Centigrade and pressure of one atmosphere, and c and b are constants.

In the last few years several rock samples have been collected to construct a data base of thermal properties of rocks for Mainland Portugal. Table 1 presents a summary of thermal conductivities of main rock types for Portugal.

**Table 1: Average thermal conductivities measured in cores from boreholes and rock samples collected in outcrops, (adapted from Ramalho and Correia, 2006; and Amaral et al., 2013).**

| Rock type | Number of cores | Number of boreholes | Average thermal conductivity (W/mK) |
|---|---|---|---|
| Schist and greywacke | 22 | 3 | 3.69 |
| Schist | 2 | 1 | 3.03 |
| Schist with manganese | 12 | 2 | 4.29 |
| Basalt | 7 | 1 | 3.31 |
| Violet tuff and Tuffy schist | 4 | 1 | 5.41 |
| Green tuff | 4 | 3 | 4.29 |
| Grey tuff | 11 | 1 | 3.65 |
| Silica tuff | 4 | 1 | 3.17 |
| Greywacke | 6 | 1 | 3.48 |
| Granite | 3 | 3 | 3.05 |
| Granite | 7 | * | 2.78 |
| Dolomitic limestone | 5 | 6 | 3.83 |
| Dolomite | 4 | 1 | 3.88 |
| Limestone | 3 | 1 | 3.42 |
| Calco-Schist | 2 | 1 | 3.56 |

* Hand sample

### 2.2 Heat Production

Knowledge of heat production in rocks is fundamental to study and model the thermal regime of the crust and lithosphere. Heat production in rocks results from the radioactive decay of $U^{235}$, $U^{238}$, $Th^{232}$, and $K^{40}$. The amount of those elements is, most of the times, determined by gamma ray spectroscopy or elemental chemical analysis and the heat production is calculated using the equation:

$$A = \rho \cdot \left(9.52 \cdot C_U + 2.56 \cdot C_{Th} + 3.48 \cdot C_K\right) \cdot 10^{-5} \tag{11}$$

where $A$ is the heat production per unit volume in $\mu W/m^3$, $\rho$ is the density in $kg/m^3$, and $C_U$ and $C_{Th}$ are the concentrations of uranium and thorium in ppm, respectively, and $C_K$ is the concentration of potassium in %.

Table 2 shows the average heat production values for some rocks of Mainland Portugal. Because most of those rocks were collected in shallow outcrops, there is an uncertainty about the concentration of the radioactive elements with depth. In those cases, it is common practice (even though there is some debate about the value of the method) the following equation is used to estimate the heat production with depth:

**Table 2: Average heat production per unit volume for some rock types from Mainland Portugal.**

| Rock type | Average heat production ($\mu W/m^3$) |
|---|---|
| Granite (ERG) | 2.80 (0.05) |
| Granite | 0.832 * |
| Microgranite (ERG) | 2.75 (0.25) |
| Syenite | 2.161 * |
| Diorite (ERG) | 1.17 (0.14) |
| Microdiorite (ERG) | 0.80 (0.07) |
| Gabbro (ERG) | 0.19 (0.02) |
| Gabbro | 0.046 * |
| Peridotite | 0.025 * |

(ERG) stands for gamma ray spectroscopy (adapted from Correia et al., 1993; Correia, 1995; Correia and Jones, 1997; Correia and Carrilho Lopes, 2008). The values in parenthesis are standard deviations. The values with * stand for elemental chemical analysis.

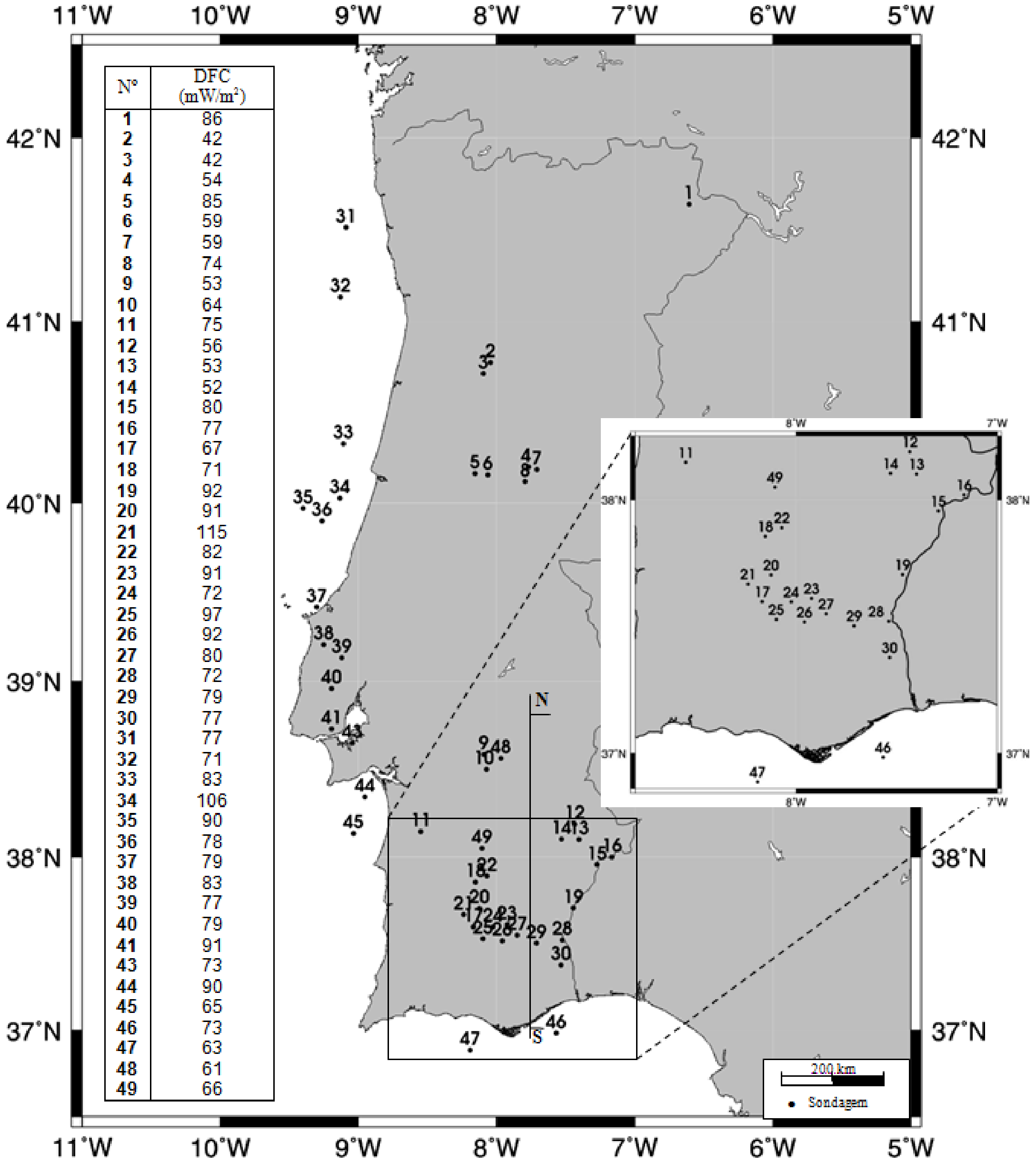
$$ln\,A = 12.6 - 2.17 \cdot V_P \qquad (12)$$

where $ln$ is the natural logarithm, $A$ is the heat production per unit volume in $\mu W/m^3$, and $V_p$ is the P wave velocity in the rock type where the heat production is to be estimated.

In the geothermal model of Figure 2, because there was no reliable heat production measurement/estimate for the deeper parts of the crust, eq. (12) was used. Details about the model in Figure 2 can be seen in Correia and Safanda (2002).

## 3. RESULTS AND DISCUSSION

Figure 1 shows a map of Mainland Portugal with the location of the boreholes used in this paper and the heat flow density values (attached table). A geothermal model and a temperature profile along the N-S profile shown in Figure 1 are shown in Figure 2 and Figure 3, respectively.




| Nº | DFC ($mW/m^2$) |
|---|---|
| 1 | 86 |
| 2 | 42 |
| 3 | 42 |
| 4 | 54 |
| 5 | 85 |
| 6 | 59 |
| 7 | 59 |
| 8 | 74 |
| 9 | 53 |
| 10 | 64 |
| 11 | 75 |
| 12 | 56 |
| 13 | 53 |
| 14 | 52 |
| 15 | 80 |
| 16 | 77 |
| 17 | 67 |
| 18 | 71 |
| 19 | 92 |
| 20 | 91 |
| 21 | 115 |
| 22 | 82 |
| 23 | 91 |
| 24 | 72 |
| 25 | 97 |
| 26 | 92 |
| 27 | 80 |
| 28 | 72 |
| 29 | 79 |
| 30 | 77 |
| 31 | 77 |
| 32 | 71 |
| 33 | 83 |
| 34 | 106 |
| 35 | 90 |
| 36 | 78 |
| 37 | 79 |
| 38 | 83 |
| 39 | 77 |
| 40 | 79 |
| 41 | 91 |
| 43 | 73 |
| 44 | 90 |
| 45 | 65 |
| 46 | 73 |
| 47 | 63 |
| 48 | 61 |
| 49 | 66 |

**Figure 1: Location (map) and heat flow density values (attached table) for the 49 boreholes of this study. The location of the profile N-S is also shown (see Figure 2 for geothermal model).**

The southern part of mainland Portugal has a higher concentration of estimated heat flow density values than the northern part. This is a result of mining activity which is more intense in the south than in the north of the country. The N-S profile line shown in Figure 1 crosses two basic geologic units called the Ossa-Morena Zone (to the north) and the South-Portuguese Zone (to the south); those zones are separated by a deep overthrust known as the Ferreira-Ficalho Overthrust (see Figure 2). These two units have different geological, tectonic, and volcanism characteristics. Using the approach described in this study it is possible to infer that

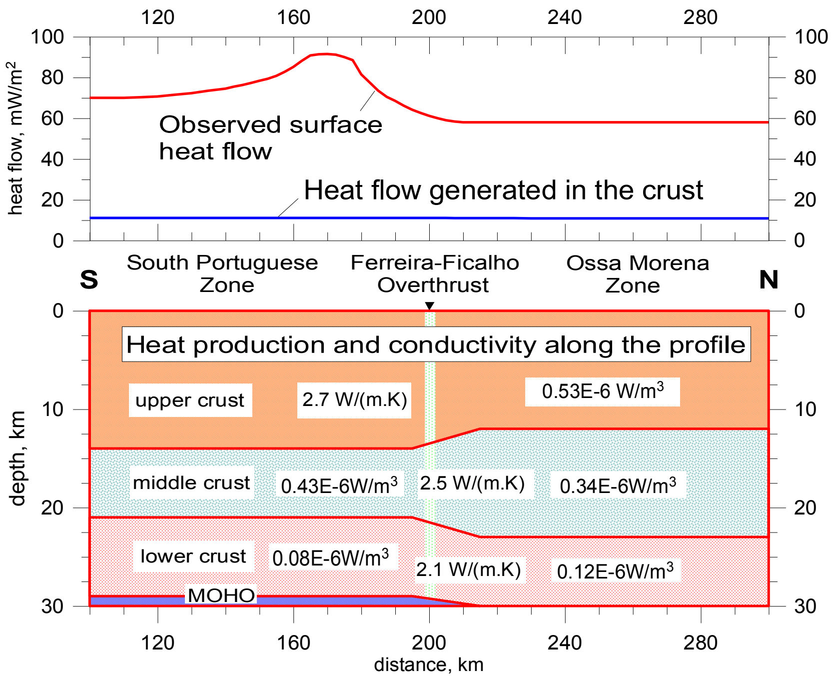



**Figure 2: Geothermal model along profile N-S shown in Figure 1. The three layered crust is inferred from seismic surveys and the thermal conductivity values are from Correia and Jones (1995) and Correia and Ramalho (1999).**

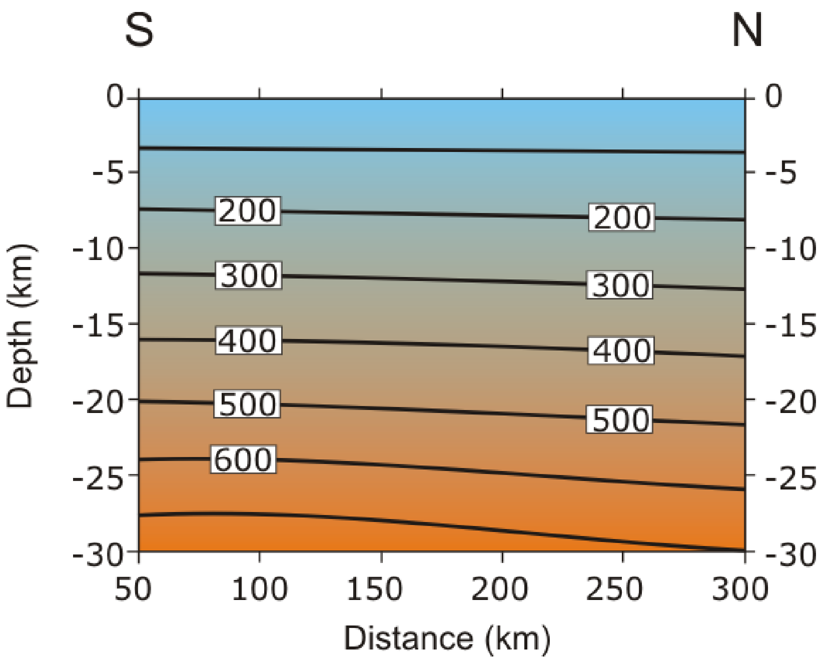



**Figure 3: Temperature distribution along profile N-S of Figure 1. The model used to calculate the temperature is the geothermal model of Figure 2. Temperature values in degrees Centigrade.**

temperature in the base of the crust should not be higher than 700 °C; however, for the Ossa-Morena Zone the temperature at the base of the crust must be lower than in the South-Portuguese Zone. An interesting result from the modeling is that, on average, the heat flow density estimated for the surface is about 20 mW/m$^2$ lower than the actual heat flow density values. One possible explanation is that the estimates for the heat production for the three crustal layers are too low (Correia and Safanda, 2002). Another feature of the model that has no correspondence with actual measurements is that near the overthrust shown in Figure 2 the regional heat production should vary between 4.0 and 4.8 μW/m$^3$; these values are, however, much higher than the heat production values estimated from cores or rock samples from the region. The high heat flow density values can, therefore, be a result of other phenomena such as thermal conductivity heterogeneities or fluid flow. Of course new temperature, thermal conductivity, and heat production data are necessary to resolve these local inconsistencies. One fact, though, is true: the heat flow density values estimated for southern Mainland Portugal are in the same range as the heat flow density values estimated for other European Hercynian regions (Cermak and Lastovickova,1987; Fernandez et al.,1998). For Mainland Portugal, the heat flow density varies between 42 and 115 mW/m$^2$.

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