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Key Points:

- Curie point depth corresponds to the 580°C isotherm calculated for the Iberian Peninsula and its margins
- Geodynamical implications of the Curie-point depth isotherm are assessed
- Heat flow map and Moho temperature map are presented

Supporting Information:

- Supporting Information S1

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Curie Point Depth of the Iberian Peninsula and Surrounding Margins. A Thermal and Tectonic Perspective of its Evolution

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Abstract In this work the thermal structure of the Iberian Peninsula is derived from magnetic data by calculating the bottom of the magnetization, assumed to be the Curie-point depth (CPD) isotherm, which accounts for the depth at which magnetite becomes paramagnetic (580°C). Comparison of the CPD with crustal thickness maps along with a heat flow map derived from the CPD provides new insights on the lithospheric thermal regime. Within Iberia, the CPD isotherm has thickness in the range of 17 to 29 km. This isotherm is shallow (<18 km) offshore, where the lithosphere is thinner. In continental Iberia, the NW Variscan domain presents a magnetic response that is most probably linked to thickening and later extension processes during the late Variscan Orogeny, which resulted in widespread crustal melting and emplacement of granites (in the Central Iberian Arc). The signature of the CPD at the Gibraltar Arc reveals a geometry consistent with the slab roll-back geodynamic model that shaped the western Mediterranean. In offshore areas, a broad extension of magnetized upper mantle is found. Serpentinization of the upper mantle, probably triggered in an extensional context, is proposed to account for the magnetic signal. The Atlantic margin presents up to 8 km of serpentinites, which, according to the identification of exhumed mantle, correlates with a hyperextended margin. The Mediterranean also presents generalized serpentinization up to 6 km in the Algerian Basin. Furthermore, a heat flow map and a Moho temperature map derived from the CPD are presented.

1. Introduction

The thermal structure of the crust and lithosphere is related to ancient and active tectonics, and its knowledge constrains geodynamic reconstructions. It provides insights on the internal state of the mantle, its evolution, and geochemistry as the mechanical/physical properties of the lithosphere are temperature-dependent (Afonso et al., 2005). Understanding of the regional variations in the thermal regime is an important target in geophysics, as it influences crustal magmatism, seismicity, rheology, the distribution of the seismic wave velocities, and seismic wave attenuation. Li and Wang (2016) state that surface heat flow variations cannot be faithfully used in inferring deep thermal structures, which can be better constrained overall through Curie-point depth (CPD) maps derived from magnetic anomalies. Furthermore, heat flow is one of the most elusive geophysical observables, it is difficult to measure directly, and its measurements are irregularly distributed and characterized by large variability (Davies, 2013). Therefore, the calculation of the CPD is a good proxy for the thermal configuration of the lithosphere.

The Curie point is the temperature above which a magnetized material becomes paramagnetic. In the Earth's crust, iron-titanium oxides (titanomagnetites) are the main source for the magnetic anomalies (e.g., O'Reilly, 1976). The temperature of the Curie point for these oxides depends on the titanium content. In acid rocks the titanium fraction is low and the Curie point can be approximated to 580°C (the value for magnetite). In some basic rocks the titanium fraction can be important and the Curie point decreases to values closer to 500°C, or even much lower values for some extrusive rocks (Herrero-Bervera et al., 2011). However, these minerals occur only in some tectonic settings, and therefore, the Curie point of magnetite (580°C) is often used as temperature at the CPD.

The topography of the CPD can be used to estimate the geothermal gradient, the heat flow, and the lithospheric thickness of an area, all of them key components in understanding the tectono-thermal structure and evolution of a region. However, the use of CPD in calculating the heat flow and in the interpretation of results relies on the assumption that it corresponds to the 580°C isotherm. Issues could arise due to

compositional heterogeneities or where the bottom of the magnetization does not correspond to the CPD because of the absence of magnetic minerals at depth.

Historically, aeromagnetic data have been used to estimate the CPD by undertaking spectral analysis in the Fourier domain. Several authors have later developed new methods to calculate the CPD, first assuming random magnetization (Bhattacharyya & Leu, 1975; Okubo et al., 1985; Tanaka et al., 1999), and later, introducing the fractal distribution of the magnetization into the equations (Bouligand et al., 2009; Li et al., 2013; Salem et al., 2014).

The aim of this work is to estimate, the bottom of the magnetization, which we assume to be the 580°C isotherm, in the Iberian Peninsula and surrounding margins. The position of this boundary provides the first step to calculate a heat flow map of Iberia. The calculated CPD map is also compared to a crustal thickness (CT) map obtained by compilation of active and passive seismic data results to assess the likelihood of an upper magnetic mantle and its tectonic implications. Lateral changes in the topography of the CPD give useful insights on the thermal state of Iberia and can be correlated with the present and past geodynamic evolution of the area. However, some of the observations raise new questions about the structure and the tectono-thermal history of Iberia.

2. Geological and Geophysical Framework

One of the geophysically best constrained areas within the European continent is the Iberian Peninsula and surrounding areas. In particular, the western Mediterranean (the Gibraltar arc and the Alboran Basin) has been the target of high-resolution multidisciplinary geophysics and geology studies since the 1980s. One of the largest research initiatives developed since 2006 has been Topo-Iberia research project, aimed to establish an integrated framework to understand the evolution of the topography within the Iberian Peninsula, viewed as the interaction between deep, shallow, and atmospheric processes (Gallart et al., 2015). A varied number of studies have targeted the lithospheric structure of Iberia. Receiver function imaging and high-resolution tomographic models have been presented (Mancilla & Diaz, 2015; Palomeras et al., 2014; Thurner et al., 2014). Studies of 2-D and 3-D thermal modeling studies were developed (Carballo, Fernández, Jiménez-Munt, et al., 2015; Carballo, Fernández, Torné, et al., 2015; Fulla et al., 2010; Torné et al., 2015). Nonetheless, studies based on large-scale magnetic data were overlooked, and thus, the thermal structure for the entire region is still loosely constrained.

The Iberian Peninsula, in West Europe (Figure 1), is geologically bounded by the Pyrenees and the Gulf of Biscay to the North and by Africa to the South. From a tectonic point of view, it can be divided in (a) Variscan Iberian Massif (IB), outcropping at its western half, forming the main Iberia basement; (b) Alpine mountain chains, including the Pyrenees, and the Iberian and Betic cordilleras; (c) the western Mediterranean area that includes the Gibraltar Arc, the Algerian basin, and the Valencia Trough and the Balearic Promontory; and (d) Atlantic domains, namely, the Gulf of Cadiz, the Western Iberian Atlantic margin, and the Bay of Biscay.

These areas are the result of different tectonic events; the Variscan Orogeny was caused by the collision of Laurussia and Gondwana in Late Paleozoic times (Matte, 2001). The IB is subdivided, from north to south, in Cantabrian Zone, West Asturian-Leonese Zone, Galicia Tras-Os-Montes Zone (GTMZ), Central Iberian Zone (CIZ), Ossa-Morena Zone (OMZ), and the South Portuguese Zone (SPZ) (Julivert et al., 1972). The Cantabrian Zone and the SPZ are interpreted as the foreland fold and thrust belts, while the rest of the areas correspond to the internal zones. The GTMZ is an allochthonous unit over the CIZ, and it is composed by Gondwana terrains and ophiolites corresponding to vestiges of oceanic crust, with high-pressure metamorphism (Martínez Catalan et al., 2014). These rocks, together with those found between the SPZ, OMZ, and CIZ, suggest the existence of one or more sutures (Simancas et al., 2013). The significance of the CIZ is currently under discussion. Some authors thought it to define the so called Central Iberian Arc (CIA) (Martínez Catalán, 2012), which, together with the Ibero-Armorican Arc, would define a double Variscan orocline. Pastor-Galán et al. (2015, 2016, 2017) argue that it is not a double orocline but, based on paleomagnetic data, a curve originated as a combination of processes occurred in Variscan times and later in the Cenozoic, during the Alpine tectonics. The Alpine domains were formed after the collision between Eurasia and Africa involving the Iberian microplate, which was disaggregated during the breakup of Pangea in the Mesozoic. As a result, Alpine chains such as the Iberian Chain were formed. Some

Figure 1. Simplified geological map of the study area with major tectonic provinces and structures. CS: Central System; CVF: Calatrava Volcanic Field, CCR: Catalan Coastal Range. Location of granitoids is taken from Simancas et al. (2013). The units from the Iberian Massif (IB) are as follows: CIZ: Central Iberian Zone, CZ: Cantabrian Zone, GTMZ: Galicia Tras-os-Montes Zone, SPZ: South Portuguese Zone, OMZ: Ossa-Morena Zone, WALZ: West Asturian-Leonese Zone. The dark shaded areas represent the basement of the Alpine chains. The bold lines represent geologic zones and major structures.

Variscan areas were reactivated and uplifted, for example, Pyrenees, Central System, and Cantabrian Mountains; the Betic-Rif orogenic system together with the Alboran Domain is a complex compressional-extensional area formed by the collision between Africa and the Iberian microplate. Different models have been proposed to explain the geodynamics of the area. There is a general agreement on slab roll-back as the primary mechanism to account for the tectonic setting. However, two main models have been proposed to explain the current tectonic configuration of the arc as well as to explain the kinematic evolution of the orogen. Rosenbaum et al. (2002) proposed a model where the Alboran domain migrated from east to west by subduction rollback. The second hypothesis (Vergès & Fernández, 2012) suggests that the Alboran Domain was originated from a SE dipping subduction of the crust from the Ligurian Tethys, and the consequent rollback toward the NW. Various geophysical studies portray a delaminating slab attached at the Betics and Rif (Palomeras et al., 2014; Thurner et al., 2014), where a thickening of the crust is observed (Díaz et al., 2016).

3. Data and Methods

3.1. Magnetic Data

The total magnetic field anomaly data (TMI) (Figure 2) are based on a compilation provided by Getech (www.getech.com) and cover the Iberian Peninsula as well as northern Africa, from 45°N to 34°N and from 10°W to 5°E. The magnetic data are included in Getech's European and Mediterranean Magnetic Project, which compiles data derived from many magnetic surveys with various acquisition parameters (Fletcher et al., 2011). The African part of the TMI map comes from Getech's African Magnetic Mapping project (Green et al., 1992). All surveys were checked for errors and corrected prior to being joined together to form the contiguous digital grid. All data were adjusted to 1 km above terrain and are provided as continuous, merged 1 km grids (details on surveys and processing methodology can be found in previous references). The EMAG2 model (Maus et al., 2009) was used to fill some gaps in the data and was resampled to 1 km grid spacing and downward continued to adjust it to 1 km height above terrain before merging it with Getech's data.

Figure 2. Map of the total magnetic field anomalies of the Iberian Peninsula and North Africa used in this study with the same nomenclature as Figure 1. The data come from Getech Group Plc. Data gaps have been filled with the EMAG 2 data set.

3.2. Crustal Thickness Map

For the purpose of comparing the bottom of the magnetization with the CT map in the study area, we have used a combination of data sets to represent the latter (Figure 3): Diaz et al. (2016) supplies most of the CT of the continental Iberian Peninsula and the Valencia Trough. For the Algerian Basin, data come from Lepître et al. (2013) and Carballo, Fernández, Torné, et al. (2015), while the global data set CRUST 1.0 (Laske et al., 2013) has been used for the areas where there is no local data (Figure 3). The data from Diaz et al. (2016) result from active and passive seismic sources but lack information in the Atlantic Domain, which is covered by the CRUST 1.0 compilation (Laske et al., 2013). The data provided by Carballo, Fernández, Torné, et al. (2015) come from potential field modeling.

3.3. Depth to the Bottom of the Magnetization

Different methodologies have been proposed to estimate the bottom of the magnetic sources (Z_b) where the magnetization of rocks is lost due to temperature (CPD), with Fourier spectral methods being the main tool. Bhattacharyya and Leu (1975) developed a method for calculating the depth to the centroid (Z_c) of a parallel-lemped assuming constant magnetization. The method was further developed by Okubo et al. (1985) and Tanaka et al. (1999) accounting for random magnetization. This method employs the logarithm of the Fourier power spectrum to calculate the depth to the bottom of an infinite magnetic layer. Assuming random magnetization, Tanaka et al. (1999) simplified the calculus of Blakely (1995) formulating the radial average power-density spectra of the total-field anomaly $\Phi_{\Delta T}$ as

$$\Phi_{\Delta T}(k) = Ae^{-2|k|Z_t} \left(1 - e^{-|k|(Z_b - Z_t)}\right)^2 \quad (1)$$

where k is the wavenumber, A is a constant, and Z_t and Z_c are the top and centroid to the magnetic layer, respectively. This equation was simplified as

Figure 3. Moho map compiled from the data sets from Diaz et al. (2016), Carballo, Fernández, Jiménez-Munt, et al. (2015), and Laske et al. (2013). The bold lines represent geologic units as seen in Figure 1.

$$\ln \left[\Phi_{\Delta T}(|k|)^{1/2} \right] = \ln B - |k| Z_t \quad (2)$$

and

$$\ln \left[\Phi_{\Delta T}(|k|)^{1/2} / |k| \right] = \ln C - |k| Z_c \quad (3)$$

Both are linear equations where B and C are constants and Z_t and Z_c are the slopes that can be calculated fitting a straight line at high and low wavenumbers in (2) and (3), respectively. Once these are calculated, Z_b can be computed as

$$Z_b = 2Z_c - Z_t \quad (4)$$

Recently, several authors have pointed out that calculation of Z_b using this technique may lead to overestimated values due to the assumption of random magnetization (Bouligand et al., 2009; Salem et al., 2014). Further developments have been made to the method assuming that the magnetic layer responds to a fractal distribution of sources (Bouligand et al., 2009; Maus & Dimri, 1996; Maus et al., 1997). Li et al. (2013) presented a modified centroid method calculation to account for a fractal response of the magnetic layer. This approach has been successfully applied in different scenarios (Bansal et al., 2011; Li et al., 2013; Li & Wang, 2016; Wang & Li, 2015) to model the depth to the bottom of the magnetic layer as follows,

$$\ln \left[|k|^{(\beta-1)/2} \times A_{\Delta T}(|k|) \right] = \ln B - |k|Z_t \quad (5)$$

$$\ln \left[|k|^{(\beta-1)/2} \times A_{\Delta T}(|k|)/|k| \right] = \ln C - |k|Z_c \quad (6)$$

where $A_{\Delta T}(|k|)$ is the radially averaged amplitude spectra, Z_t is the top of the centroid, Z_c is the centroid depth, B and C are constants, and $A_{\Delta T}(|k|)/|k|$ is the scaled amplitude spectrum.

Z_t and Z_c are estimated from equations (5) and (6), respectively, fitting a straight line in the high and low wavenumbers, and Z_b is calculated from equation (4).

The fractal magnetization parameter β is strongly correlated to Z_t and Z_c making the simultaneous inversion of the three parameters very complicated (Bansal et al., 2011; Bouligand et al., 2009). Thus, fixing the fractal exponent provides a better and stable solution.

Blakely (1995) stated that the base of the magnetization could be captured if the window size is big enough as to account for the longest wavelengths. The response of the magnetization is a peak in the spectrum where the central wavenumber of the peak is related to the depth of the bottom of the sources. Salem et al. (2014) used this assumption to correct the power spectrum of the magnetic data in order to detect the peak in every window. For this study we have performed the calculation of Z_b with the modified formulation of the centroid method by Li et al. (2013). As stated by Ravat et al. (2016) the selection of the wavenumber range where the slope is picked has to be checked manually for each window and cannot be automated. Therefore, we have checked each window and selected the appropriate wavenumber range to calculate the bottom and the centroid of the magnetic layer. The wavenumber selection is based on the group of points after the spectral peak that gives the greater slope of the spectrum. Furthermore, we have adopted the methodology of Salem et al. (2014) in order to better constrain the fractal parameter. Our approach was based in visual inspection of the corrected amplitude spectrum applying different fractal parameters. We have modeled the amplitude spectrum with the calculated values of Z_b and Z_t to check the goodness of fit and a misfit map is provided (Figure S4.)

3.3.1. Window Size

The window size is a critical parameter in the bottom of the magnetization calculation as it constrains the wavelength that will be captured in the window. Several authors (Blakely, 1995; Bouligand et al., 2009; Maus et al., 1997) have studied the range of acceptable window sizes to better estimate the depth to the bottom of the magnetization. Blakely (1995) pointed out that the window size should be around five times the expected CPD, while a study of Ravat et al. (2007) suggested that the window size should be at least 200–300 km. Increasing the size of the window presents various issues as (a) decrease of resolution of the resulting map, (b) there is a big chance of mixing different tectonic environments, making the estimation of the bottom of the magnetization more complicated. However, increasing the window size ensures the capture of the longest wavelengths.

After reviewing the literature, we have performed tests using different window sizes in order to evaluate its impact on the results. Three different window sizes have been tested in three locations at different tectonic settings: 150×150 km, 200×200 km, and 300×300 km. Results show that the CPD differences between the smallest and the biggest window are around 10%. We have selected these window sizes taking into account (1) the smallest window size where a peak is found at low wavenumbers, and (2) in the study area, the Moho is expected to be around 30 km depth in continental areas, increasing in the Pyrenees, and shallowing to offshore areas (<20 km). Therefore, and for the sake of resolution, we have adopted two different window sizes, 150×150 km² and 200×200 km², with a shift between windows of 50 and 66.66 km, respectively. We have calculated a total of 940 windows and taken the averaged CPD, derived from the two window sizes selected, as our final solution as done by Wang and Li (2015).

3.3.2. Fractal Exponent

Several authors have studied the implications of using different fractal exponents. Bouligand et al. (2009) suggested that a value of $2 \leq \beta \leq 3.25$ depending on the tectonic setting is appropriate. Bansal et al. (2013) pointed out that a fractal exponent above 1.5 might overcorrect the spectrum resulting in shallower estimations. For this study, we have compared the outcome of several fractal exponents using visual

Figure 4. (a) Example of the change in the shape of the original amplitude spectrum (red line) after correction for different fractal exponents. Note the displacement of the peak from small wavenumbers to higher wavenumbers when increasing the exponent. In orange, modeled spectrum using the calculated depth to bottom and depth to top values. (b) Example of a final corrected spectrum (black line) and the scaled spectrum (red dashed line) with selected segments to calculate Z_t and Z_c .

inspection of the behavior of the amplitude spectrum as described by Salem et al. (2014). We tested values of $\beta = 2, 2.5, 3, 3.5$, and 4 (Figure 4a) for every window and compared the resulting de-fractal spectra. We have later selected one among the previous exponents for each window attempting to correct the fractal distribution (Figure 4b). The selection of β was accomplished by assuming that the best value was the one showing a peak at small wavenumbers with at least two points at its left-hand side. Furthermore, as β increases, the spectrum is overcorrected, which could lead, in some cases, to the displacement of the peak to higher wavenumbers, thus evidencing the invalidity of that specific parameter.

Some authors (Ravat et al., 2016; Salem et al., 2014) have used the centroid method along with forward modeling of the spectral peak (Blakely, 1995; Connard et al., 1983) to constrain β . Although the comparison could give reliable results, it depends on the calculated values of Z_b and Z_t , and thus in the selection of the wavenumber range where the above parameters are calculated, making the comparison dependent on the subjectivity of picking wavenumber ranges.

Our solution leads to a high variability in β values (Figure S3). For marine areas such as the Alboran Sea or the oceanic crust present in the Algerian Basin, β needs to be small, between 2.5 and 3 matching the results from previous studies (Li et al., 2013; Li & Wang, 2016; Wang & Li, 2015). For continental areas β varies between 3 and 4 due to the more heterogeneous geological setting of the continental Iberian Peninsula. In general, $\beta = 3$ is the best fit for the S and N-NE of the Iberian Plate whereas $\beta = 3.5$ works for the central and central-west part. In some particular cases, NW of continental Iberia, we had to choose $\beta = 4$ to correctly account for the fractal distribution of sources.

Figure 5. Curie-point depth (CPD) map of the Iberian Peninsula averaged from the CPD maps calculated with window sizes of $150 \times 150 \text{ km}^2$ and $200 \times 200 \text{ km}^2$. The map shows the thickness of the magnetic layer. The isolines correspond to contours of the CPD. The bold lines represent geologic units as seen in Figure 1.

4. Results

4.1. Curie Point Depth

CPD corresponds to the depth at which rocks in a particular location reach their Curie temperature. It is assumed that this is an isotherm and corresponds to the bottom of the magnetization calculated by spectral analysis of magnetic data. This property provides the link between magnetism and thermal state (in the lithosphere). This surface represents the base of the magnetization below topography onshore and below bathymetry offshore. Therefore, it can be regarded as the thickness of the layer that is magnetized. For clarity, the CPD will be described and discussed in depth below elevation (topography and bathymetry). In the study area of the Iberian Peninsula the CPD obtained with the approach presented in this paper corresponds to a surface located at depths between 27 and 13 km below elevation (Figure 5). The resulting CPD map shows clear differences between east and west, and also between onshore and offshore areas. Considering CPD as an isotherm, we can define hot areas where CPD is shallow and cold areas where CPD is deep. Cold values characterize the older and more stable Iberian Massif and the Alpine domains to the east, including the Cenozoic basins. Hot areas lie mostly offshore.

The West Mediterranean domain, including the Alboran Sea, is characterized by relatively shallow CPD, ranging between 17 and 20 km below elevation. There are three minima in this area: (1) the Alboran Sea, (2) the Algerian Basin, and (3) north of the Balearic Promontory. All three describe a SW-NE lineation subparallel to the Mediterranean Coast.

These shallow values reach the interior of the Iberian Peninsula mostly in the NE (Figure 5). A prominent feature is the relatively deep (~3 km deeper), N-S CPD located in the Valencia trough. The southern part of the study area, which includes the Gibraltar Arc and N Africa, is represented by deep CPD values reaching depths between 22 to 24 km. Here the geometry of the Curie isotherm follows the geometry of the Betic-Rif orogenic system in the continental areas and continues to the Tell Mountains in NE Africa. To the south, in the edge of the study area, the CPD becomes shallower. Westward of the Gulf of Cadiz, the Atlantic margin shows shallow values ranging from 21 up to 16 km in its westernmost end, minimum values that extend over the Western Iberia Atlantic Margin, with the exception of the region offshore in the same latitude of the Galicia Bank. These values extend to the south of Portugal but do not reach the interior of the Peninsula, being bounded by the Gibraltar Arc deep CPD values. In central Spain, shallow CPD values (up to 22 km) appear locally over the Quaternary Calatrava Volcanic Field (CVF). To the NW of Iberia, the CPD deepens reaching more than 26 km. These deep values extend to the east, getting to the Central System and part of the Iberian Chain. Over the Pyrenees there is not a marked trend in the CPD values, being quite variable and locally shallow, in contrast with the deeper values (22–24 km) observed in the western region of the Cantabrian Mountains. The Bay of Biscay presents shallow CPD (<20 km) toward the north except locally near -7° . Marked minima of the CPD within the continental areas are located to the NE beneath the Catalan Coastal Ranges, in the SW area of Portugal, and in the western limit of the Algerian basin.

A global model of the CPD was presented by Li et al. (2017). In this work the estimation of the CPD was performed using the EMAG2 data set, and calculating the CPD with a constant fractal exponent of 3. Comparison of these results with our CPD calculation presents consistent results. The mean value of the misfit between both maps is 3.4 km with the Li et al. (2017) results deeper than the results presented here. The minimum disagreement is found in the Mediterranean area where the misfit is mainly between ± 2 km. Maximum differences are found in the Gibraltar Arc and in the Iberian Chain. There, a sharp contrast is found in the data set of Li et al. (2017) from the Mediterranean to the onshore. The contrast is presented as a straight change with N-S direction. This sharp feature in the Li et al. (2017) results could represent a problem in the data set used to calculate the CPD or an incorrect selection of β for the area.

4.2. Comparison of Crustal Thickness With Curie Point Depths

The relationship between the CT (Figure 3) and CPD values (Figure 5) is important to assess whether the mantle is magnetic or not and thus, to establish the tectonic settings where the mantle may contribute to magnetic anomalies. The CPD/CT offset (Figure 6) depicts negative values representing CPD values deeper than the Moho and positive values representing areas where the CPD is shallower than the Moho. Onshore, the CPDs are mainly above the Moho (up to 25 km) and offshore is locally below this boundary (up to 10 km). Continental and thermally stable areas like the Iberian Peninsula are expected to feature CPDs above the Moho, since they are thermally reequilibrated regions at present.

Among the onshore areas, the Betic-Rif and Pyrenees mountain belts present the highest differences between the CT and CPD as deep crustal roots have been proposed for both orogens (Díaz et al., 2016). In addition, the CPD/CT offset in the Pyrenees extends to the south into the Iberian Chain. The western part of the Iberian Peninsula, mostly corresponding to the Iberian Massif, is more homogeneous presenting CPD/CT differences around 8 to 10 km depth. The NW part of the Iberian Massif shows an elongated, NNW-SEE directed area where the CPD/CT differences are below 8 km, even descending to 4 km in its central part.

For offshore areas we find that the CPD is usually located beneath the Moho, such as in the Atlantic Margin areas, including the Gulf of Cadiz and the Bay of Biscay, the Alboran Sea, and the Algerian Basin. The mean depth difference between CPD and Crustal thickness in these areas is 3 km but locally can get up to 8 km. The CPD/CT offset characterizing the western Mediterranean is high in part of the Alboran Sea and increases constantly toward the Algerian Basin, and the Valencia Trough, with a local decrease in the Balearic Promontory. This configuration indicates an altered upper mantle.

However, not every offshore area presents CPD values deeper than the Moho. The Atlantic margin does feature CPD deeper than the Moho along the coast line. But in the Gulf of Cadiz, this negative offset is displaced to the WSW where the Moho map (Figure 3) places this boundary at depths within the 20–25 km range. The

Figure 6. Map showing the differences between the crustal thickness (Figure 3) and the Curie point depth (CPD) (Figure 5). Negative values (red) correspond with CPD deeper than the Moho (black dashed line). Positive values (orange/green/blue) represent CPD above the Moho. The bold lines represent geologic units as seen in Figure 1.

Bay of Biscay is also characterized by a Moho deeper than the CPD close to the coast but this configuration seems to change to the N.

4.3. Heat Flow Map

We have calculated a heat flow (Q_0) map (Figure 8) for the Iberian Peninsula and surrounding margins from the CPD map assuming steady state condition. The surface heat flow for a layer has two components: the background heat and the heat production inside the layer from radiogenic decay. From the Fourier's law, the 1-D surface heat flow in steady state, with constant heat production (A), constant thermal conductivity (k), surface temperature (T_0), and temperature at Z depth (T_z), is as simple as

$$Q_0 = A \cdot Z / 2 + (T_z - T_0)k / Z \quad (7a)$$

In our case, $T_z = T_{cdp}$ is the temperature of the bottom of the magnetic layer. When CPD is above the Moho, we calculate as a one-layer problem:

$$Q_0 = A \cdot Z_{cpd} / 2 + (T_{cpd} - T_0)k / Z_{cpd} \quad (7b)$$

but when CDP is below the Moho, we calculate as a two-layer problem (Figure 7):

$$Q_0 = \frac{[T_{cpd} - T_0 + A \cdot Z_m \cdot (CPD - Z_m) / km + A \cdot Z_m^2 / 2kc]}{[Z_m / kc + (CPD - Z_m) / km]} \quad (8)$$

where Z_m is Moho depth and kc and km are respectively the thermal conductivity of the crust and the mantle.

Figure 7. Surface heat flow has been calculated as a problem of one layer when the Moho is deeper than Curie point depth (CPD) and as a problem of two layers when CDP is deeper than the Moho. We assume that heat production (A) and thermal conductivity (k_c for the crust and k_m for the mantle) are constant.

Values of surface temperature for continental Iberia were taken from a map of mean surface temperature from Hijmans et al. (2005). For offshore areas, the source of information was taken from Boyer et al. (2005). The calculations have been performed assuming steady state conditions. Values for thermal parameters such as thermal conductivity and radiogenic crustal heat production are assumed constant for the crust and upper mantle. Thermal conductivity of 2.2 W/mK is used for the crust where $CT > CPD$, as a mean value accounting for sediments, upper and lower crust. In offshore areas where the CPD is thicker than the CT, a value of 2.7 W/mK has been used to account for the altered upper mantle (Osako et al., 2010) in the Mediterranean area. We assumed a radiogenic heat production of $1 \mu\text{W m}^{-3}$ for the crust (FFernández et al., 1998) and null for the lithospheric mantle.

The Q_0 map presents similar features to the CPD map shown in Figure 5. In general, continental areas present the lowest values ranging from 60 to 72 mW/m² except for some coastal regions where values up to 80 mW/m² are reached.

Low Q_0 is found throughout the IB, with minima in the GTMZ area. We believe that this minimum derives from the use of constant thermal parameters for the Q_0 calculations as this is a highly crystalline area where high thermal conductivity values are expected. Relative minima are found as well in the Betics and in the Rif. Offshore areas present high Q_0 , with the Atlantic margin showing the highest values. The Mediterranean portrays a NE-SW lineation of high Q_0 , except in the Balearic Promontory where a minimum of 72–76 mW/m² is found. Singularities in this map are found over the Tell Mountains, where a relative maximum stands out over the relative low values of the eastern Atlas Mountains. In continental Iberia, two singularities are found: the relative maxima enclosing at the CVF in central Iberia and that at the Sierra de la Demanda, to the North.

A comparison with available heat flow (Q_0) measurements (Fernández et al., 1998) depicts a good correlation between the calculated Q_0 and the measured Q_0 values (Figure 8) despite that short wavelength anomalies cannot be captured by the gradient map due to its resolution. Small discrepancies could be regarded as the effect of using constant values of thermal conductivity and radiogenic heat production. The best correlation is found in the western Mediterranean where most data are concentrated. Transitions from high to low Q_0 measured values are captured by the calculated Q_0 map in the Algerian Basin and toward the west in the Alboran Sea and Gibraltar Arc. In onshore areas, the lack of a good Q_0 value coverage makes the relationship between both variables difficult. However, good correlations are found in the SW Iberian Peninsula, where high Q_0 coincides with the measured values. Still, direct comparison between both variables should be taken with care. The calculated Q_0 has a better coverage onshore than the measured one. Furthermore, Q_0 data respond to the thermal state of the shallowest crust and can be affected by hydrothermal processes, which may alter the measurements.

4.4. Moho Temperature Map

Following the calculation of the Q_0 map for the Iberian Peninsula and surrounding areas, we have computed the temperature at the Moho (Figure 9). We have used the same thermal parameters as in the calculations of Q_0 , according to the steady state formulation,

$$T_m = -A \cdot Z_m / 2k_c + Q_0 \cdot Z_m / k_c + T_o \quad (9)$$

where T_m is the temperature of the Moho, Z_m is the Moho depth, k_c is the crustal thermal conductivity, and T_o is the surface temperature.

The map portrays a clear differentiation between Variscan and Alpine terrains, where the latter are clearly hotter than the former. The maximum values are found where the crust is thicker, Pyrenees and Rif, where

Figure 8. Heat flow map of the Iberian Peninsula and surrounding areas derived from the Curie-point depth map presented in Figure 5. The bold lines represent geologic units as seen in Figure 1. The white lines represent heat flow tendencies measured by Fernández et al. (1998).

the temperature is between 800 and 850°C. For the rest of continental Iberia, the Moho temperature values are in the range of 600–750°C, with the exception of SW Portugal where values of 750–800°C are found. For offshore areas, values are in the range of 600–400°C being lower in the Algerian Basin and close to the Alboran Sea in the Mediterranean and in the Atlantic.

5. Discussion

Magnetic data of the Iberian Peninsula and surrounding margins are used to estimate the bottom of the magnetic layer, which we reconcile with the CPD. Assuming that magnetization responds to magnetite content, the resulting CPD represents the 580°C isotherm. This is later compared with a CT map of the Iberian Peninsula (Figure 6) in order to provide constraints in its thermal regime. Furthermore, the data are then used to compute the Q_0 map and Moho temperature map for the area (Figures 8 and 9). The more relevant findings and implications of this study are addressed in the following sections.

5.1. Betic-Rif Orogenic System/Gibraltar Arc

The Gibraltar arc is described as an arcuate orogenic system originating from the convergence of the Iberian microplate with the African plate. Different geophysical studies have pointed out the possibility of the Alboran delaminating slab to be attached at some parts to the orogen (Palomeras et al., 2014; Thurner et al., 2014). From a thermal perspective, the presence of a sunken slab would perturb the thermal regime, decreasing the geothermal gradient (Syracuse et al., 2010) and thus deepening the CPD and decreasing Q_0 . These issues are consistent with the maps provided in Figures 5 and 7. We therefore suggest that deep CPD/low Q_0 values in the Gibraltar Arc and Betic-Rif orogen are consistent with a thickened crust (55 km in

Figure 9. Moho temperature map of the Iberian Peninsula and surrounding areas derived from the Curie-point depth map presented in Figure 5. The bold lines represent geologic units as seen in Figure 1.

the Rif; Díaz et al., 2016) and lithosphere. If this relationship is correct, it can be proposed that the slab is still attached or has been recently detached under the Betics between 6°W and 3°W, and 37°N and 38°N. According to our data, the slab might have a NE-SW direction with a significant change in the trend of the CPD shallowing toward the east. In the African margin, deep CPD and low Q_0 are observed in the Rif area (in northern Morocco) continuing to Algeria into the Tell Mountains. There, the E-W trend of the CPD changes its orientation to NE-SW, probably in relation to the crustal thickening of the latter mountain belt, continuation of the Atlas Mountains.

To the south of the Rif Mountains, the CPD becomes shallower as we approach the Atlas Mountains. Even though the Tell Mountains are a prolongation of the Atlas Mountains and the same CPD distributions should be expected, a lithospheric thinning has been identified underneath the latter (e.g., Teixell et al., 2005; Zeyen et al., 2005; Palomeras et al., 2014) possibly related to the Canary Islands mantle plume (Duggen et al., 2009) or lateral flow of the Alboran slab (Levander et al., 2014). Our data support the existence of a high thermal gradient in the Middle and High Atlas even though this area is located at the southern edge of our data set. This may lead to shallower CPD values (Figure 5) and high Moho temperatures (Figure 9) in response to the asthenospheric upwelling and is also consistent with outcrops of Quaternary volcanism.

5.2. Western Mediterranean

The western Mediterranean tectonic evolution is strongly linked to the development of the Gibraltar Arc. Geographically, it includes the Alboran domain, the Algerian Basin, the Valencia Trough, and the Balearic Promontory (Figure 1).

As seen in the CT map, the western Mediterranean represents an area of thin crust (~15 km). Comparison of the CT with our final solution for the base of the magnetized layer (Figure 6) reflects that the Moho is shallower than the CPD, suggesting an almost generalized magnetization of the upper mantle in the Mediterranean. The magnetization follows the geometry of the Iberia and Africa shorelines and reaches onshore areas in the Tell Mountains and locally in NE Iberia. The magnetized layer extends northward until the Valencia Trough, where no magnetic mantle is identified near the coast. Instead, contours are displaced to the east, until they reach the Ibiza Island within the Balearic Promontory, where mantle seems to contribute to the magnetization again. The maps suggest a relatively thick layer of magnetized mantle in the Algerian Basin and the east part of the Alboran domain, but none close to the subduction zone. Another thick layer of magnetized mantle is found in the Valencia Trough, located between the coast and the Balearic Islands.

Magnetic mantle is commonly interpreted as an alteration of peridotites to serpentines (Ferré et al., 2013). In serpentines, magnetite is formed as a secondary phase, and it is responsible for most of the magnetic signal within the mantle. Serpentinites appear in some geodynamic contexts, such as subduction zones or extended margins. In the former, they result as the product of the mantle alteration due to hydration in the trench and the slab wedge, forming a thin layer that favors the exhumation of high-pressure rocks. In the latter, they appear as an interaction between extension, exhumation, and hydration. The change in mineralogy that serpentization implies disturbs the rheological properties of the mantle (Guillot et al., 2015) due to the significant reduction in density and strength involved. This provokes a change in buoyancy that has been interpreted as a mechanism that further triggers crustal extension (Guillot et al., 2015). In this context, we propose that the magnetized mantle present in the western Mediterranean is formed by intrusion of water in the system, creating the alteration of upper mantle to serpentinites. The CPD/CT offset in the western Mediterranean (Figure 6) could represent the maximum thickness of serpentinites. However, the possibility that the CPD just pictured the bottom of serpentinites, and not the depth to the 580°C isotherm, cannot be excluded. The latter would imply that serpentinites might exist deeper, but above the T_c (580°C), and therefore, the heat flow deduced would be biased in this area.

Previous studies have pointed out the presence of altered mantle in the West Mediterranean. Different authors have studied the Valencia Trough and found slow velocities in the upper mantle from wide-angle and refraction seismic sections (Torné et al., 1992) and from thermal modeling (Carballo, Fernández, Torné, et al., 2015). Our study portrays a relatively thick layer of serpentinite of variable thickness, ranging from 1 up to 6 km in the center of the Valencia Trough. Furthermore, serpentinitized peridotites have been exhumed in the Alpujárride Complex (southern Spain) and in Beni Bousera (northern Morocco) (Pedrera et al., 2016). Figure 6 also shows magnetic mantle in the Algerian Basin. It is reasonable to correlate these features with serpentization since the Algerian Basin is likely formed by oceanic crust.

The geodynamic evolution of the western Mediterranean is highly complex, and the origin of the serpentinitized upper mantle could be attributed to different causes and temporal scenarios. First, it could be the result of the extension due to slab roll-back of the Gibraltar Arc. Extensional tectonic settings favor the intrusion of water through normal faulting (Guillot et al., 2015). The increasing buoyancy induced by the serpentization could severely feedback in the amount of crustal extension. This may be the case of the thick serpentinitized layers in the Valencia Trough and Algerian Basin, although the first was aborted during Late Miocene (Etheve et al., 2016), while the second progressed. Thus, the serpentization could be the expression of the upper mantle evolution during the postslab retreat.

However, the presence of serpentines might not only be the result of the most recent geodynamic evolution of the Mediterranean, but an inherited feature. A process like the subduction rollback in the West Mediterranean needs a low strength interface to trigger the decoupling of the slab. Thus, the presence of a layer of hydrated mantle could act as a weak shear zone. The Balearic Promontory underwent different extensional events associated with the opening of the Alpine-Maghrebian Tethys from Late Triassic to Early Cretaceous, and extensional-compressional events from Late Cretaceous to Cenozoic. This evolution is similar to that of adjacent areas such as the West Alps, Corsica, and the Pyrenees, where hyperextended rifted margins and exhumed serpentinites have been described (Beltrando et al., 2014; Jammes et al., 2009). Some authors have made numerical modeling of the evolution of the subduction rollback (Chertova et al., 2014). Their results pictured the initiation of the lithospheric decoupling within the Valencia Trough

Figure 10. (a) S wave velocity of the upper mantle as a function of degree of serpentinization. The red dots represent laboratory measurements by Christensen (1966); the grey dots are based on V_s velocities from Palomeras et al. (2017). (b) Degree of serpentinization of the upper mantle. The extent is derived from the Curie-point depth/crustal thickness offset, and the percentage is obtained from Figure 10a.

and the Balearic Promontory, between the Iberian coast, Ibiza, and Mallorca. The location of this feature coincides with a relative maximum in our CPD map and the major thickness of magnetized upper mantle in the area. Accordingly, the presence of a serpentinite layer in this point could have helped the decoupling of the lithosphere and the initiation of the slab retreat. As a result, the current serpentinization of the West Mediterranean could be the expression of different stages of its geodynamic evolution, covering the pre-syn-subduction and postsubduction rollback range. The altered peridotites could be consequence of hydration of the mantle wedge, extension due to slab roll-back, or a combination of both. We propose that serpentinization of the mantle in the West Mediterranean might have played a crucial role in the geodynamics of the area and could be responsible of the initiation of the lithospheric detachment, exhumation of high-pressure rocks, and the uplift of the Betic-Rif orogen due to the reduction of the density and buoyancy of the mantle. The age of serpentinization cannot be deduced from this work as it can be linked to several tectonic stages.

Serpentinization produces a decrease in the seismic velocities of the mantle due to hydrated mineral assemblages. In order to assess the degree of serpentinization in the upper mantle, we have used S wave velocity information from Palomeras et al. (2017) as proposed by Christensen (1966). As the S wave information available is restricted to the Mediterranean, data were set to a pressure of 400 MPa appropriate for the base of an ~15 km thick crust. For a high degree of serpentinization (96%), the S wave velocity could be as low as 2.29 km/s, whereas velocities of 4.6 km/s correspond to fresh mantle (Christensen, 1966). To compute the velocity of the magnetized Mediterranean mantle, we have averaged the S wave velocities of the layer comprised between the Moho and the CPD. These velocities (Figure 10a) allow us to estimate the mean percentage of serpentinization of the magnetized upper mantle by comparing them with laboratory measurements. Results presented in Figure 10b show the percentage of hydrothermal alteration of the mantle of the western Mediterranean. We can observe a variable degree of serpentinization ranging from as little as 10%, increasing up to 33% to the E-NE presenting a maximum in the Algerian Basin. The Valencia Trough presents low values of serpentinization with maxima of 15%. The percentage of serpentinization does not show a linear relationship with the thickness of magnetized mantle, and thick magnetic mantle does not always present major degrees of serpentinization. The Algerian Basin presents the highest percentage and the largest thickness of the magnetic layer (Figure 10b), which complies well with the evolution of an area

Figure 11. Curie-point depth map of the Iberian Peninsula overlain with outcropping Variscan granites (white polygons) and the Alboran Domain volcanism (dashed white polygons). The bold lines represent geologic units as seen in Figure 1.

that was affected by intense extensional processes and thought to be composed by oceanic crust. As discussed above, the origin of serpentinization could respond to different mechanisms and times, thus hindering any further interpretations.

5.3. Variscan Belt

The thickness of the magnetic layer shows quite a homogeneous response throughout the Variscan Belt, where the CPD is located at least 22 km below the topography. A prominent feature is the elongated small CPD/CT offset at the NW Iberian Peninsula. This area corresponds to the GTMZ and the NW part of the CI2, extending south to the Central System and the Tajo Basin. This feature coincides with the position of the internal part of the CIA (Martínez Catalán, 2012; Martínez Catalan et al., 2014). The emplacement of GTMZ allochthonous complex took place during the contractional events of the Variscan Orogeny (Martínez Catalan et al., 2014) along large thrust systems. This stage represented the pressure peak during the collisional event and resulted in a thickened crust of up to 60 km (Martínez Catalan et al., 2014). This thickening led to heating and extension, and the area coincides with the larger volume of extruded granites within the Iberian Massif. Two principal groups have been described (Bea, 2004): one consisting of synkinematic granodiorites derived from lower crustal melting, with ages of approximately 325–320 Ma, and another derived from partial melting of the middle crust with age approximately 320–305 Ma. Actually, Ayarza et al. (2016) suggests that the magnetic anomalies defining the CIA all coincide with gneiss domes and probably have the same origin. There is a straightforward correlation (Figure 11) between the area occupied by outcropping and occasionally magnetic granites, the geometry of this minimum CPD/CT offset, and the area described by Martínez Catalan et al. (2014) based on the development of gneiss domes. Accordingly, we suggest that the oval shape of the minimum corresponds to an area of similar Variscan tectonic evolution

characterized by the emplacement of the allochthonous complexes over the CIZ, which resulted in Variscan crustal thickening, melting, and extension. These processes are responsible for the present magnetic CPD signature of this area. Thermal relaxation must have followed this stage, and no posterior significant thermal events are identified. To the south of the Iberian Massif, the south of the CIZ does not present the same CPD signature limiting the extension of the processes described above. Furthermore, we propose that this cold anomaly may be a consequence of high thermal conductivity linked to a highly crystalline crust in this area. The western part of the OMZ and SPZ presents the shallowest CPD values for the entire Variscan area. We suggest that they respond to the existence of generally basic igneous rocks emplaced at shallow crustal levels instead of to large volumes of granites (see Figures 1 and 2 for comparison of the geology and wavelength of magnetic anomalies). Also, the presence of a near plate boundary between Africa and Iberia may play an important role on the thermal configuration of SW Iberia as well as in the Gibraltar Arc. This result agrees with high heat flow observations and has been explained by a thin lithosphere in the area (Fernández et al., 1998).

Isolated features are recognized throughout the IB. A single relative maximum is highlighted over the CVF. This could be related to the upwelling of the asthenosphere in this volcanic region and coincides with areas of thin lithosphere as pictured by *S* wave tomography (Palomeras et al., 2017). Similarly, a thermal anomaly can be identified north of the Atlas, also coincident with Quaternary volcanism and high asthenosphere (Palomeras et al., 2014; Teixell et al., 2005; Zeyen et al., 2005).

5.4. Alpine Tectonics Domains and Atlantic Margin

Alpine terranes within the Iberian Peninsula present the highest positive offsets in CPD/CT, along with the Rif domain. The Pyrenees and the Iberian Chain show constant values >12 km. The *Qo* map shows for these areas higher values than for the IB and with high lateral variability. Contrary to what might be expected, the Pyrenees shows no thermal anomaly. The reason could be that the low expected basal heat flow is compensated for by a high radiogenic heat production in the thick crust, resulting in a flat thermal signal. However, the lack of a characteristic magnetic signature in these Alpine domains, that is, magnetic anomalies in these chains are weak and sparse, provides little constraints to image the CPD.

The Atlantic domains show negative CPD/CT offsets. The direct implication is that there is a general magnetization of the upper mantle. This has been already described for the oceanic domain by Li et al. (2013), who recall extensive magnetization in oceanic mantle. Serpentinities have been described along the Iberian Margin from drilling and geophysical studies (Boillot et al., 1980; Reston, 2009). The proposed interpretation of the CPD/CT offset agrees with previous studies and could be the result of the magma-poor rifting configuration, which takes place in hyperextended crust and mantle exhumation (Reston, 2009). The CPD/CT offset map in the Bay of Biscay reveals negative values close to the thrust system, which we reconcile with serpentinites formed in a hyperthinned crust by entrance of water through the rifting process, which to the north is replaced by exhumed mantle (Tugend et al., 2014). Also, southward subduction of the oceanic crust of the Bay of Biscay (Ayarza et al., 2004) must have played an important role in the serpentinitization of the mantle wedge.

Our results are based on a methodology that uses magnetic data to compute the bottom of the magnetization. Differences in resolution within the data set could affect the results. Furthermore, we are assuming a fractal distribution of the magnetization, which is more reasonable than assuming random distribution (Salem et al., 2014), but implies the calculation of the β correction parameter. Limitations in the applicability of the method and its assumptions imply errors in the geometry of the calculated heat flow and Moho temperature map that affect the conclusions of this work. However, the coherency of the results, which are in accordance with known geological data, suggests that in the conclusions of this work are reasonable.

6. Conclusions

This is the first attempt to determine the CPD (580°C) surface (isotherm) the heat flow map and the Moho temperature map of the Iberian Peninsula with the aim of providing a view of its evolution and thermal structure. The method we have used relies on spectral analysis of magnetic data and averages the results of using two window sizes to minimize errors and ensure the capabilities to image deep magnetic sources. The resulting CPD map has been analyzed jointly with a CT map with the aim of constraining a thermal gradient of the study area.

The results (CPD and CPD/CT offset) reveal a clear differentiation between Variscan and Alpine domains. The Iberian Massif provides a more homogeneous response. An extensive area in the NW that correlates with the CIA presents deep CPD values that suggest that the Variscan evolution of the latter is linked to magnetizing processes. This area, the most internal part of the CIA, is characterized by Variscan thickening followed by intensive heating and extension in the last orogenic stages. As a consequence, these processes result in the development of an important volume of granites. A link between this highly crystalline crust and a high thermal conductivity is proposed to account for the cool anomaly of the area.

The maps offshore indicate that the West Mediterranean upper mantle is magnetic. It most probably contains areas of heterogeneous serpentinization (remnants of a hydrothermally altered mantle). This serpentinization is compatible with the geodynamics of the western Mediterranean (subduction and subduction roll-back of the Alboran lithosphere and extension of the Mediterranean basins). The serpentinites could act as the detachment level for the slab roll-back mechanism. The slab geometry in the Gibraltar Arc is captured by the signature of the CPD. In this area, the CPD increases, which can be explained by the existence of a thickened lithosphere that alters the thermal regime.

The maps reveal that, along the Iberian Atlantic margin, extensive mantle magnetization/serpentinization must have occurred, correlating with the existence of the hyperextended crust, which must have favored the exhumation of the mantle toward the W.

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