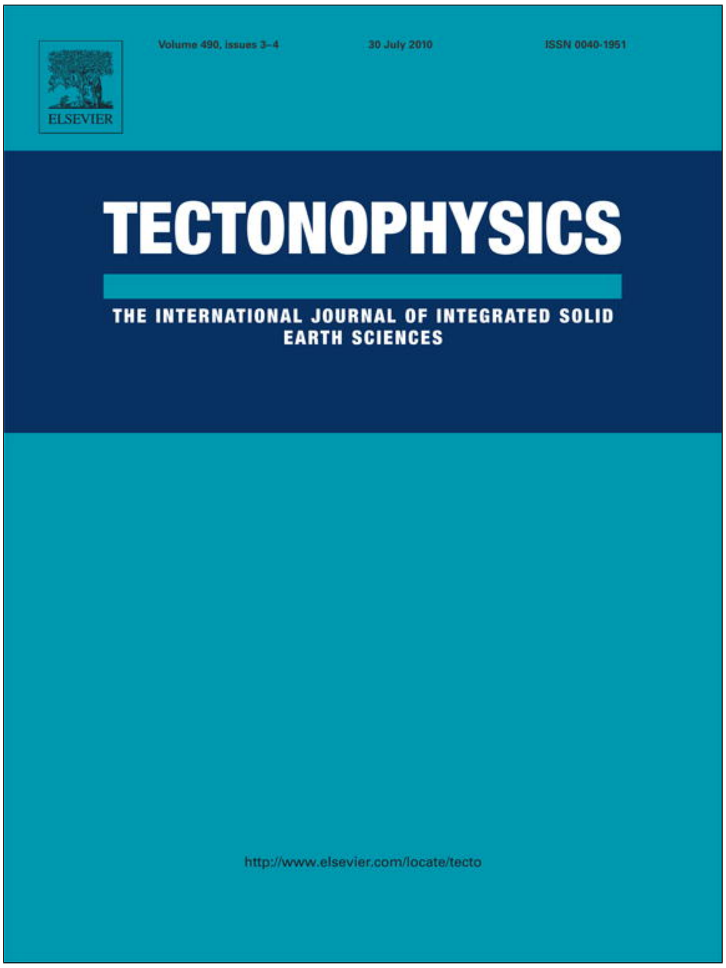
Volume 490, issues 3–4
30 July 2010
ISSN 0040-1951
ELSEVIER
TECTONOPHYSICS
THE INTERNATIONAL JOURNAL OF INTEGRATED SOLID EARTH SCIENCES
http://www.elsevier.com/locate/tecto

Tectonophysics 490 (2010) 152–164

Contents lists available at ScienceDirect

Tectonophysics

journal homepage: www.elsevier.com/locate/tecto

# Radiogenic heat production variability of some common lithological groups and its significance to lithospheric thermal modeling

M. Vilà [a,*], M. Fernández [b], I. Jiménez-Munt [b]

[a] Institut Geològic de Catalunya, Balmes, 209-211, E-08006 Barcelona, Spain
[b] Institut de Ciències de la Terra Jaume Almera (CSIC), Lluís Solé i Sabarís s/n, E-08028 Barcelona, Spain

ARTICLE INFO

*Article history:*
Received 23 July 2009
Received in revised form 12 May 2010
Accepted 12 May 2010
Available online 20 May 2010

*Keywords:*
Radiogenic heat production
Geothermal modeling
Heat flow
Lithosphere
Continental crust

ABSTRACT

Determining the temperature distribution within the lithosphere requires the knowledge of the radiogenic heat production (RHP) distribution within the crust and the lithospheric mantle. RHP of crustal rocks varies considerably at different scales as a result of the petrogenetic processes responsible for their formation and therefore RHP depends on the considered lithologies. In this work we address RHP variability of some common lithological groups from a compilation of a total of 2188 representative U, Th and K concentrations of different worldwide rock types derived from 102 published studies. To optimize the use of the generated RHP database we have classified and renamed the rock-type denominations of the original works following a petrologic classification scheme with a hierarchical structure. The RHP data of each lithological group is presented in cumulative distribution plots, and we report a table with the mean, the standard deviation, the minimum and maximum values, and the significant percentiles of these lithological groups. We discuss the reported RHP distribution for the different igneous, sedimentary and metamorphic lithological groups from a petrogenetic viewpoint and give some useful guidelines to assign RHP values to lithospheric thermal modeling.



## 1. Introduction

The distribution of the radiogenic heat producing elements within the lithosphere provides an important control on the crustal temperature distribution and the mechanical strength of the lithosphere (Sandiford and McLaren; 2006). Combined determinations of surface heat flow and radiogenic heat production (RHP) yield basic information about the thermal field and the structure of the Earth's lithosphere (Jaupart and Mareschal, 2003, 2007; Taylor and McLennnan, 1995; Beardsmore and Cull, 2001).

One of the most difficult and critical tasks in geothermal modeling is the assignment of reliable RHP values to the different geologic units that compose the lithosphere. Commonly the RHP distributions of the different layers that compose the lithosphere are assigned according to RHP determinations of representative rock samples from or near the region of interest. In many cases the inaccessibility of most of the lithospheric rocks difficults its RHP characterization. In this case published mean RHP values of representative rock types (e.g. Haack, 1983; Rybach, 1976, 1986; Wollenberg and Smith; 1987), crustal type sections (Rudnick and Fountain, 1995), geological provinces (Lewis et al., 2003; Brady et al., 2006; Perry et al., 2006) or global geothermal studies (Artemieva and Mooney, 2001; Jaupart and Mareschal; 2003) are used. However, regional studies cover only part of the crustal lithologies and, in most cases, lack of a correlation between RHP values and main lithotypes. Finally, regional geochemical studies often provide abundances of radioelements but not RHP values. In consequence, choosing average RHP values for the different lithological groups involves large uncertainties.

In this context we present a new compilation of RHP values distributed according to the main lithosphere rock types and we discuss the variability of RHP from a petrogenetic point of view. The data presented here have been obtained after revision of an extensive compilation of large geochemical and geophysical datasets previously published. The main aim of this work is to provide a representative set of RHP values and the necessary criteria for their correct use and interpretation particularly addressed to lithosphere thermal modeling.

## 2. Radiogenic elements and heat production

During the radioactive decay mass is converted to energy and an important part of this energy is converted to heat. All natural radioactive isotopes generate heat to a certain extent but only the contributions of the decay series of $^{238}U$, $^{235}U$, $^{232}Th$ and the isotope $^{40}K$ are geologically significant. The radiogenic heat production of a given rock (RHP, in μW m$^{-3}$) is calculated by (Rybach, 1988):

$$RHP = 10^{-5}\rho(9.52C_U + 2.56C_{Th} + 3.48C_K) \quad (1)$$

* Corresponding author.
*E-mail addresses:* mvila@igc.cat (M. Vilà), mfernandez@ictja.csic.es (M. Fernández), ivone@ictja.csic.es (I. Jiménez-Munt).

0040-1951/$ – see front matter. Crown Copyright © 2010 Published by Elsevier B.V. All rights reserved.
doi:10.1016/j.tecto.2010.05.003

where $\rho$ is the density of the rock (in kg m$^{-3}$) and $C_U$, $C_{Th}$ and $C_K$ are the concentration of uranium (in ppm), thorium (in ppm) and potassium (%), respectively.

Conventional geochemistry considers K as a major element (those which predominate in any rock analysis) and U and Th as trace elements (those which are present at less than 0.1% level). Assuming that at global scale the Th/U ratio is around 4 and the K/U ratio is around $1\times10^4$ (Rudnick et al., 1998; McLennan, 2001; Jaupart and Mareschal, 2003), it results that RHP is controlled basically by the U and Th concentrations (85% of the total RHP) and to a lesser extent to the K concentration.

In the case of rock samples, radiogenic heat producing elements are determined by several laboratory analytical techniques: inductively coupled plasma atomic emission spectrometry and mass spectrometry, instrumental neutron activation analysis, X-ray fluorescence spectrometry and gamma-ray spectrometry (Rybach, 1988; Darnley, 1995). Portable, car-borne and air-borne gamma-ray spectrometry techniques are used for estimating the concentrations of radioelements of the surface rocks, and borehole gamma-ray spectrometry is used to acquire a continuous spectra of the U, Th and K concentrations of subsurface rocks (IAEA, 2003).

Early work on continental heat flow observed a linear relationship between surface heat flow and near surface heat production. For zero surface heat production this linear relationship predicts a basal heat flow value, which defines the so-called Heat Flow Provinces (Birch et al., 1968; Roy et al., 1968; and Lachenbruch, 1968). This observation suggested that the variation of surface heat flow within a given heat flow province is related to variations in the total crustal heat production. Three commonly used depth distribution models of RHP that satisfy the linear relationship are the step, the linear and the exponential models (Morgan et al., 1987). From these, the exponentially decreasing model has been commonly favored because it better explains the effect of erosion (Lachenbruch, 1968). Further studies however showed that when heterogeneous lithologies are considered the linear relationship between surface heat flow and surface heat production is not satisfied and simple depth distribution models of RHP do not apply (e.g., Furlong and Chapman, 1987; Fountain et al., 1987; Kukkonen and Lahtinen, 2001). Studies of exposed crustal sections suggest a general trend of decreasing heat production with depth, but on a scale of a few kilometers there is no simple stratigraphic control on the distribution of RHP (Sandiford and McLaren, 2006; Jaupart and Mareschal, 2003; Brady et al., 2006). Rybach and Buntebarth (1982) found an experimental relationship between RHP and the cation packing index for some igneous rocks. Density and seismic velocity are directly related to the cation packing index and consequently they proposed relationships between density, seismic velocity and RHP. Since acidic rocks usually have lower densities than more basic rocks (Johnson and Olhoelf, 1982), it follows that RHP and $SiO_2$ content must be closely related. However, these relationships do not apply to all rock types and different works show that there is no straightforward relationship between RHP and seismic velocity, density and $SiO_2$ content (e.g., Fountain, 1985; Kukkonen and Peltoniemi, 1998; Joeleht and Kukkonen, 1998; Jaupart and Mareschal, 2003). Thus, presently there is no any empiric explanation that supports a simple and universal depth-dependent function for RHP distribution. Actually, several lines of evidence suggest that the ordering of the RHP within the lithosphere is related to the petrogenesis of the different geologic units that compose the lithosphere.

The large-scale distribution of the radiogenic heat production elements (RHPE) is determined by the geochemical differentiation of the Earth. In the simplest view, the primordial Earth began as an undifferentiated mass. Early in its history, the iron and related siderophilic elements sank gravitationally to the centre of the planet to form the core (32% of the mass) and giving the differentiated Silicate Earth (68% of the mass), which is also called primitive mantle and comprises the present day crust (about 0.7% of the mass of the primitive mantle) and the mantle. U, Th and K are lithophile elements and hence, they are concentrated in the Silicate Earth. On the basis of geochemical studies, the Earth's core cannot contain a significant fraction of the RHPE. Consequently, the composition of the primitive mantle is critical to our knowledge of the RHP distribution through time. The various global geochemical models of the Earth consider that initially the Silicate Earth had an homogeneous composition (e.g., Taylor and McLennan, 1985; Hofmann, 1988; Anderson, 1989; McDonough and Sun, 1995; Palme and O'Neill, 2003).

Taking $4\times10^{24}$ kg and $9\times10^{20}$ m$^3$ as the mass and the volume of the primitive mantle, and using the concentration of radiogenic elements of the present primitive mantle (K = 0.024 wt.%, Th = 0.0795 ppm and U = 0.02 ppm) proposed by McDonough and Sun (1995), we obtain that the present radiogenic heat generated by the Silicate Earth is 18 TW. If we consider that the total heat loss through the Earth's surface is 44.2 TW (Pollack et al., 1993), then the present radioactive heat generated by the Silicate Earth is equivalent to the 41% of the total heat loss through the Earth's surface.

The rate of decay of a radioactive isotope is proportional to the concentration present at any time. The concentration $C$ of a radioactive isotope at time $t$ measured backward from the present is related to the present concentration $C_0$ and to the half-life of the isotope $\tau_{1/2}$ by:

$$C = C_0 \exp\left(t \ln 2 / \tau_{1/2}\right) \tag{2}$$

Fig. 1a shows the concentration of radiogenic isotopes and elements of the primitive mantle through time proposed by McDonough and Sun (1995). It can be seen that the U concentration in the Silicate Earth at 4.5 Gyr ago was more than twice the present concentration; the initial Th concentration was only 1.25 greater than present and the K concentration experienced a very low decrease. Thus, the higher levels of U, Th and K during the early history of the Earth resulted in a higher RHP (Fig. 1b). The implication is that Precambrian, especially Archean, geotherms may have been relatively high.

U, Th and K are incompatible with respect to normal mantle minerals. During mantle partial melting and crustal formation RHPE are preferentially concentrated in the liquid phase. Consequently, RHP is strongly partitioned into the crust and the residual mantle (only those portions which participate in the differentiation process become depleted). The distribution of the RHP whithin the crust is broadly related to the distribution of lithotypes. In the case of the oceanic crust, which can be roughly subdivided into a thin sedimentary cover and a basaltic layer of nearly constant composition, is not very difficult to know its RHP distribution. In contrast from a petrological point of view the continental crust is more complex since its composition is more variable laterally and in depth (e.g., Jaupart and Mareschal, 2003; Rudnick and Fountain, 1995; Wedepohl, 1995).

In most crustal lithologies U, Th and K form part of minerals. In igneous and metamorphic rocks, K generally occurs in major silicates such as feldspars and micas, whereas U and Th contents of major minerals (quartz, feldspars, amphiboles, pyroxenes and micas) are usually negligible (Webb et al., 1987). U (as $U^{4+}$) and Th (as $Th^{4+}$) occur mainly in accessory minerals (those that are present at less than 5% level) such as zircon, monazite, xenotime, apatite, sphene and allanite (Marchalland and Fairbridge, 1999; Van Schmus, 1995). U and Th have a similar electron configuration (for Th $z = 90$ and for U $z = 92$) and similar radii ($U^{4+} = 1.05$ Å, $Th^{4+} = 1.1$ Å) and the two elements can substitute each other explaining their relative geochemical coherence although the mobility of U is much higher. Metamorphic rocks devolatilize with progressive metamorphism (Ague, 1991), hence metamorphic fluid flow could remobilize heat producing elements, particularly U, by dehydration reactions and fluid flow. In surface environment U and Th-bearing minerals may be resistant to weathering and commonly occur as detrital grains. However, $U^{4+}$ can be

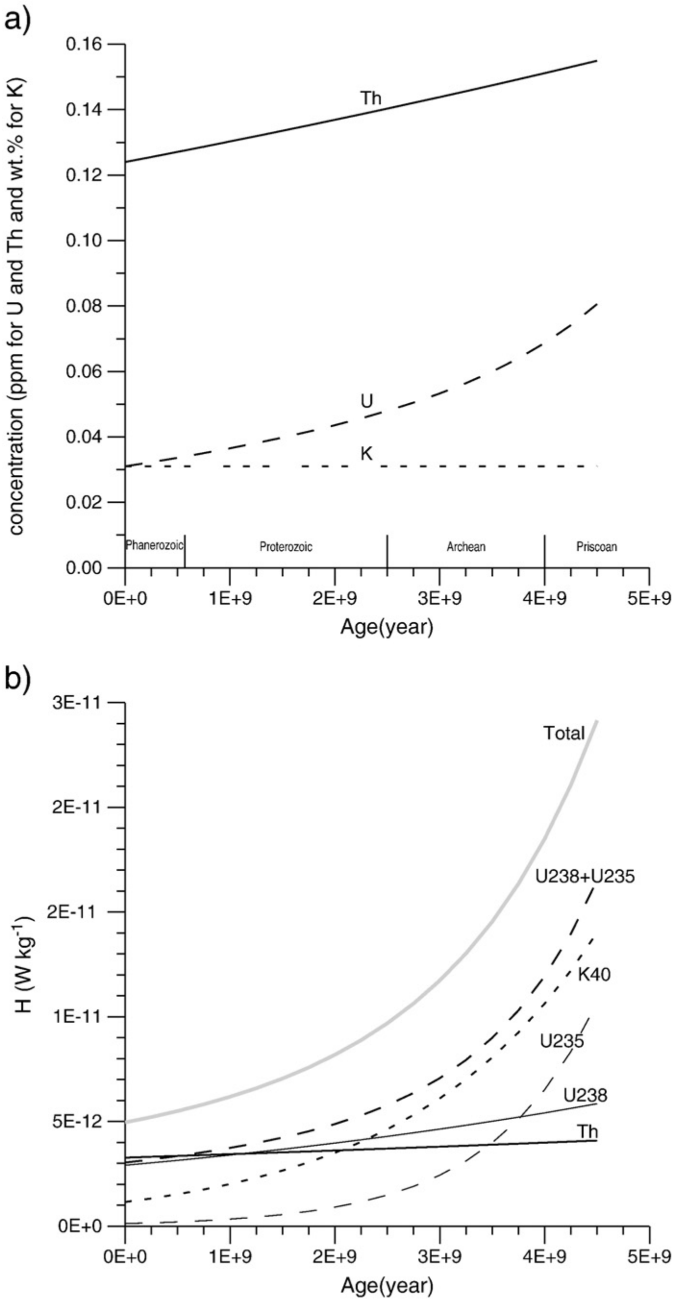


**Fig. 1.** Heat production through time assuming the silicate earth composition (primitive mantle) of McDonough and Sun (1995). a) U, Th and K concentrations due to the decay of radioactive isotopes as functions of time measured back from the present and (b) radiogenic heat production within the Earth during the past.

**Table 1**
Petrologic scheme with hierarchical structure showing the selected lithological groups. Bold: lithological groups with RHP values established from a statistical point of view. Italic: lithological groups with orientative RHP values based only on a detailed review of all compiled values.

| Earth material |
|---|
| **Igneous rock** |
| **Plutonic rock** |
| **Ultramafic** |
| *Peridotite* |
| *Pyroxenite* |
| *Hornblendite* |
| **Gabbroid–dioritoid–anorthosite** |
| **Gabbroid** (gabbro, monzogabbro) |
| **Dioritoid** (diorite, monzodiorite) |
| *Anorthosite* |
| *Syenitoid* |
| **Granitoid** |
| **Tonalite** (trondhjemite) |
| **Granodiorite** |
| **Granite** (monzogranite,syenogranite) |
| *Alkali feldspar granite* |
| *Foid syenitoid* |
| *Foid dioritoid–foid gabbroid* |
| *Foidolite* |
| **Volcanic rock** |
| **Basalt** |
| *Basaltic andesite* |
| **Andesite** |
| *Dacite* |
| **Rhyolite** |
| *Trachyte* |
| *Phonolite* |
| *Tephrite–basanite* |
| *Foidite* |
| **Sedimentary rock** (consolidated and unconsolidated) |
| **Detritic** |
| **Mudrock** (shale, claystone, siltstone, argillite, lutite, mud, muddy sediment) |
| **Wacke** (graywacke) |
| **Sandstone** (arenite, sand, sandy sediment) |
| *Quartzarenite* |
| *Arkose* |
| *Litharenite* |
| *Conglomerate* |
| **Carbonate** |
| *Limestone* |
| *Dolostone* |
| *Marl* |
| *Organic-rich* |
| *Phosphate-rich* |
| *Evaporite* |
| *Non-detritic siliceous* |
| **Metamorphic rock** |
| **Low grade–medium grade** (up to amphibolite facies) |
| **Metaigneous** (gneiss, amphibolite, green schist) |
| **Metasedimentary** (slate, phyllite, schist, gneiss, pelite, metasandstone, marble) |
| **High grade** (high-amphibolite facies, granulite facies, eclogite facies) |
| **Granulite** |
| *Felsic granulite* |
| *Intermediate granulite* |
| *Mafic granulite* |
| **Eclogite** |

oxidized to $U^{6+}$, which is more mobile (Chamberlain and Sonder, 1990; Plant et al., 1999). Weathering of feldspars and micas favours the formation of K-bearing clay minerals such as illite.

## 3. Heat production and lithotypes

The main objective of the present study is to obtain a table of representative RHP values of the main lithological groups that compose the lithosphere. To achieve this goal, we have created a database that contains 2188 representative RHPE concentrations of different rock types derived from 102 published studies. The different studies provide a median number of 16 samples and the 90% of them provide less than 50 samples. A complete table with all the data used and a list of references is included in the Appendix.

The compiled samples correspond to the major tectonic plates and include the main geodynamic settings of the Earth. The samples cover an age interval ranging from the Archean to Recent. The compiled RHPE concentration data basically result from single samples collected from outcrops and boreholes that we assume as pertaining to essentially uniform lithologies. Only 499 RHPE concentrations come from composite samples of which 196 come from previously published global average compositions of different rock types. The rest of the composite samples represent average concentrations of local and regional geologic units of roughly homogeneous lithologies. RHP has been computed from the U, Th and K concentrations using the formula (1) proposed by

Rybach (1988). A reasonable average density, based on the studies of Johnson and Olhoelf (1982) and Christensen and Mooney (1995) was assigned for each lithologic group.

To build a comprehensible RHP database focused on thermal modeling we have classified and renamed samples following a petrologic classification scheme with a hierarchical structure (Table 1). Establishing appropriate lithological groups is a complex and hard task because of the heterogeneity of the available petrological information. Many of the works used do not include appropriate petrological descriptions or the classification schemes used to name the samples. In addition, the geological parameters used to describe a rock sample (e.g. color index, mineral content, grain size mode, chemical composition, age, geological setting, facies, etc.) vary from study to study. Therefore, similar rocks are denoted with different names due to a continued use of traditional terms or without justified reasons. This reflects that, despite the great number of petrologic compendiums, a standard and widely accepted scheme to define and classify rocks is still lacking.

Initially, the earth materials forming the lithosphere have been grouped in three main lithological groups corresponding to igneous, sedimentary, and metamorphic rocks. We have subdivided the igneous rocks into plutonic and volcanic varieties according to the criteria proposed by the IUGS (Le Maitre, 2003). Sedimentary rocks are firstly classified into six categories based on sediment composition, without considering the degree of consolidation. In a lower hierarchical level, sedimentary detritic rocks are classified according to the grain size. The primary classification level of metamorphic rocks is based on the metamorphic grade. Low grade and medium grade metamorphic rocks have been subdivided into metaigneous and metasedimentary varieties. Within the high-grade metamorphic rocks we have distinguished granulites and eclogites.

The proposed hierarchical structure allows the user to classify the wide range of rock types in a logical and unambiguous manner and provides a simple way for inputing, storing and retrieving data. In addition, the classification and naming of rocks can be varied according to the expertise of the user and the level of available information (the more information is available, the higher is the level of the hierarchy at which the rock can be classified).

## 4. Results

To assign statistically significant RHP values to the different lithological groups we first considered those groups with more than 20 values derived from more than 4 representative studies. For the rest of lithological groups, we propose estimated RHP values after a detailed review of all the reported values and bearing in mind the classification relationships between the different lithological groups. In order to detect potential outliers inside each lithological group, those maximum values with distances greater than 50% of the nearest value were investigated. Eventually, some samples were excluded after a careful analysis of the reported values and the cumulative plots. Afterwards, the statistical results have been compared with RHP values obtained from previously published U–Th–K average concentrations for the different lithological groups.

Table 2 summarizes the standard deviation, the minimum and maximum values, and the 10th, 25th, 50th, 75th and 90th percentiles of the different lithological groups; and Fig. 2 shows the RHP cumulative distribution plots of the three main lithological groups, igneous rocks, sedimentary rocks and metamorphic rocks.

Despite the outlier exclusion, the highest RHP values show a prominent increase in the slope of the cumulative plots (specially those higher than the 90th percentile) and then decrease the reliability of the mean and the standard deviation of the lithological groups. For this reason we find preferable to use the median value (50th percentile) and the range of values defined by the 25th and 75th percentiles, the

**Table 2**
Statistics of the RHP values compiled for the different lithological groups.

| Lithological group | n.val. | Density | Mean | Std. dv. | Min. | 10th percentile | 25th percentile | 50th percentile | 75th percentile | 90th percentile | Max. |
|---|---|---|---|---|---|---|---|---|---|---|---|
| earth material | 2188 | 2750 | 1.588 | 1.752 | <0.001 | 0.159 | 0.504 | 1.181 | 2.076 | 3.394 | 24.685 |
| igneous | 1218 | 2750 | 1.755 | 2.068 | <0.001 | 0.104 | 0.453 | 1.212 | 2.437 | 3.761 | 24.685 |
| igneous/plutonic | 861 | 2750 | 2.081 | 2.271 | <0.001 | 0.194 | 0.677 | 1.695 | 2.686 | 4.191 | 24.685 |
| igneous/plutonic/ultramafic | 62 | 3300 | 0.226 | 0.232 | <0.001 | 0.002 | 0.031 | 0.111 | 0.438 | 0.548 | 0.810 |
| igneous/plutonic/gabbroid–dioritoid–anorthosite | 153 | 2900 | 0.535 | 0.540 | 0.010 | 0.056 | 0.160 | 0.371 | 0.809 | 1.230 | 3.845 |
| igneous/plutonic/gabbroid–dioritoid–anorthosite/gabbroid | 67 | 2950 | 0.468 | 0.370 | 0.052 | 0.082 | 0.179 | 0.345 | 0.722 | 0.998 | 1.467 |
| igneous/plutonic/gabbroid–dioritoid–anorthosite/dioritoid | 43 | 2850 | 0.981 | 0.686 | 0.252 | 0.355 | 0.551 | 0.862 | 1.199 | 1.632 | 3.845 |
| igneous/plutonic/granitoid | 583 | 2700 | 2.520 | 2.155 | 0.087 | 0.761 | 1.342 | 2.079 | 3.009 | 4.649 | 24.685 |
| igneous/plutonic/granitoid/tonalite | 54 | 2750 | 1.784 | 1.861 | 0.087 | 0.363 | 0.611 | 1.043 | 2.195 | 3.901 | 8.496 |
| igneous/plutonic/granitoid/granodiorite | 92 | 2700 | 2.073 | 1.235 | 0.471 | 0.762 | 1.438 | 1.879 | 2.511 | 3.135 | 8.011 |
| igneous/plutonic/granitoid/granite | 309 | 2650 | 2.827 | 2.176 | 0.263 | 1.004 | 1.741 | 2.429 | 3.233 | 4.930 | 24.685 |
| igneous/volcanic | 357 | 2650 | 0.970 | 1.133 | 0.009 | 0.067 | 0.196 | 0.573 | 1.211 | 2.568 | 6.554 |
| igneous/volcanic/basalt | 166 | 2750 | 0.358 | 0.394 | 0.009 | 0.036 | 0.077 | 0.214 | 0.533 | 0.837 | 1.899 |
| igneous/volcanic/andesite | 49 | 2650 | 0.781 | 0.362 | 0.067 | 0.294 | 0.472 | 0.818 | 1.094 | 1.212 | 1.416 |
| igneous/volcanic/rhyolite | 56 | 2550 | 2.671 | 1.385 | 0.291 | 0.650 | 1.847 | 2.551 | 3.433 | 4.826 | 6.307 |
| sedimentary | 464 | 2400 | 1.100 | 0.675 | 0.014 | 0.318 | 0.605 | 1.055 | 1.510 | 1.806 | 5.901 |
| sedimentary/detritic | 395 | 2400 | 1.189 | 0.660 | 0.032 | 0.450 | 0.705 | 1.163 | 1.579 | 1.827 | 5.901 |
| sedimentary/detritic/mudrock | 214 | 2400 | 1.392 | 0.702 | 0.165 | 0.574 | 0.968 | 1.442 | 1.657 | 1.864 | 5.901 |
| sedimentary/detritic/wacke | 34 | 2400 | 0.984 | 0.535 | 0.125 | 0.318 | 0.548 | 0.993 | 1.212 | 1.806 | 2.184 |
| sedimentary/detritic/sandstone | 128 | 2400 | 0.896 | 0.468 | 0.032 | 0.327 | 0.536 | 0.819 | 1.206 | 1.603 | 2.055 |
| sedimentary/carbonate | 32 | 2400 | 0.477 | 0.356 | 0.064 | 0.091 | 0.216 | 0.416 | 0.618 | 0.916 | 1.752 |
| metamorphic | 506 | 2800 | 1.634 | 1.512 | 0.019 | 0.215 | 0.524 | 1.288 | 2.274 | 3.206 | 12.652 |
| metamorphic/low grade–medium grade | 327 | 2750 | 1.715 | 1.473 | 0.023 | 0.238 | 0.619 | 1.490 | 2.390 | 3.195 | 12.652 |
| metamorphic/low grade–medium grade/metaigneous | 111 | 2750 | 1.334 | 1.841 | 0.023 | 0.092 | 0.242 | 0.529 | 2.024 | 2.997 | 12.652 |
| metamorphic/low grade–medium grade/metasedimentary | 130 | 2750 | 1.990 | 1.011 | 0.108 | 0.690 | 1.210 | 1.842 | 2.765 | 3.128 | 5.098 |
| metamorphic/high grade | 179 | 2900 | 1.486 | 1.575 | 0.019 | 0.093 | 0.458 | 1.026 | 1.967 | 3.367 | 10.635 |
| metamorphic/high grade/granulite | 83 | 2900 | 1.461 | 1.945 | 0.019 | 0.093 | 0.352 | 0.753 | 1.613 | 3.911 | 10.635 |
| metamorphic/high grade/eclogite | 21 | 3300 | 0.258 | 0.216 | 0.024 | 0.024 | 0.060 | 0.247 | 0.435 | 0.659 | 0.731 |

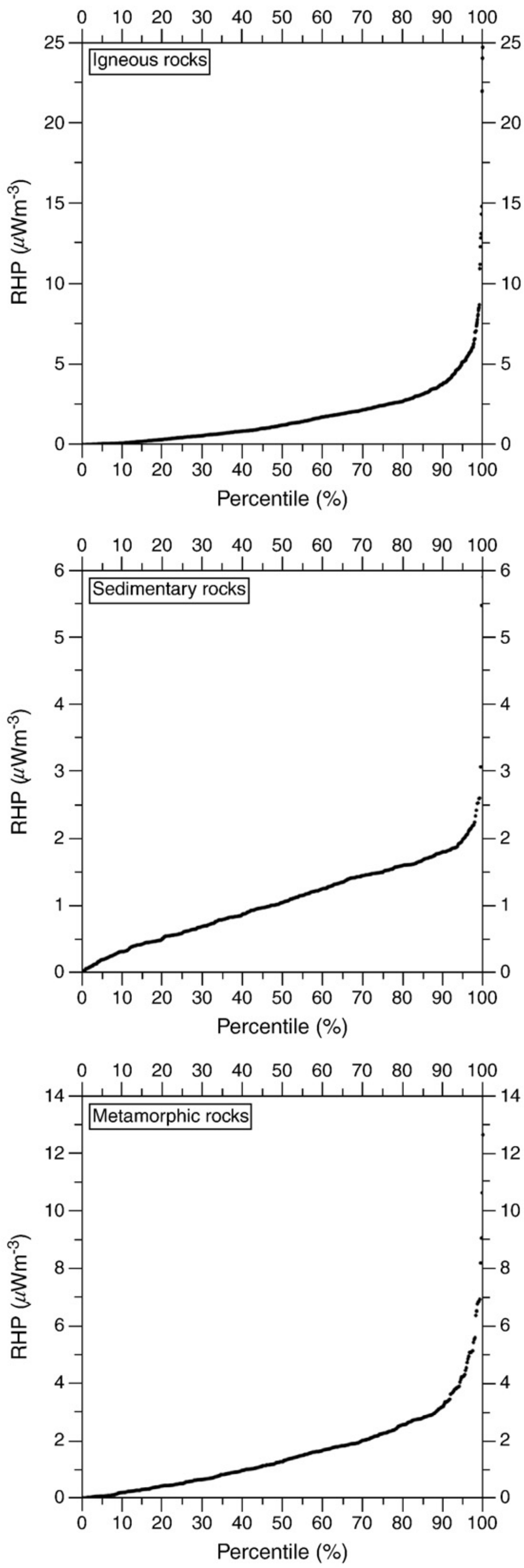


**Fig. 2.** RHP cumulative distribution plots of the main lithological groups.

10th and 90th percentiles and the minimum and the maximum values to discuss the RHP distribution within and among the different lithological groups. A clear example of this is in Fig. 2a, where the highest RHP values for igneous rocks, particularly those higher than 95th percentile, strongly influence the mean value (1.59 μW m$^{-3}$), which is 0.41 μW m$^{-3}$ higher than the median value (1.18 μW m$^{-3}$).

The cumulative curves in Fig. 2 show that for each main lithological group there is a wide zone around the median RHP value with a constant slope indicating RHP values with the same probability of occurrence. This zone usually includes the RHP range defined by the 25th and the 75th percentile and in some cases extends to values below the 10th percentile and above the 90th percentile.

A common feature of RHP distribution is that there is a considerable overlap between the values reported for the different lithological groups, even within the same geological region. Part of this overlap is because the main lithological groups are necessarily very generic and they include a large number of specific lithologies. Nonetheless, specific lithotypes pertaining to the same geological region often display a RHP variability comparable to that of the main lithological groups from that they belong. Among other main reasons to be considered are: (a) U and Th are trace elements and therefore their concentrations are not determinant variables to classify rocks from a geologic point of view, (b) the boundary lines in any classification scheme do not denote naturally occurring divisions, and (c) a rock classification imposes arbitrary subdivisions in a continuum medium, and consequently, part of the variation of the RHP assigned to a lithotype is due to these subdivisions. Finally, rock names have different meanings depending upon the used classification scheme and in many cases the assignment of names to rocks is based on a vague and diffuse petrologic characterization.

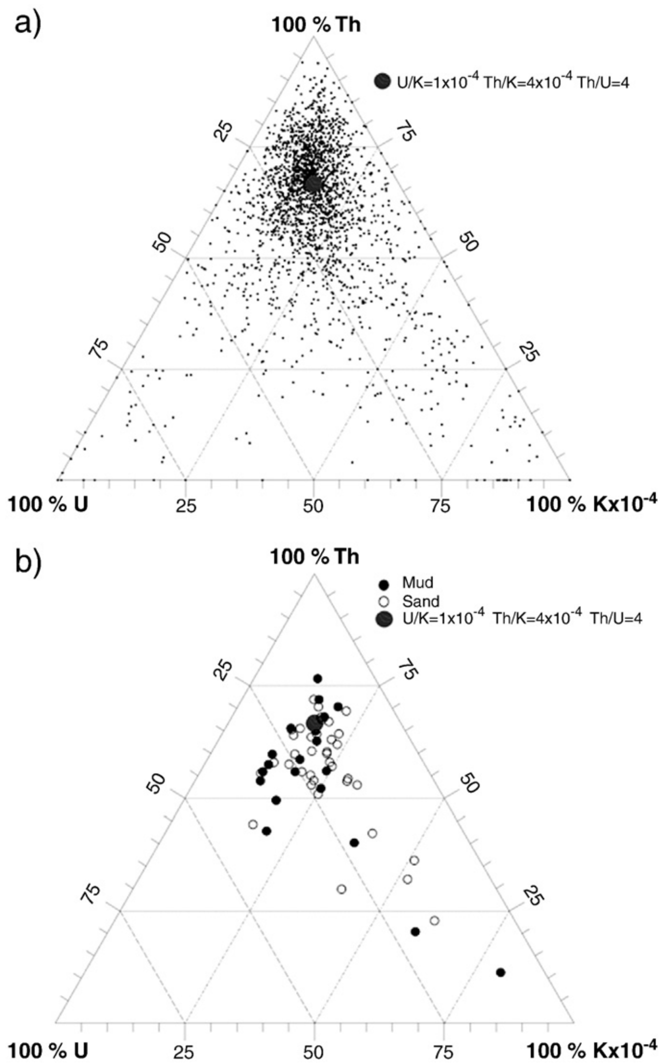


**Fig. 3.** a) Triangular projection of the U–Th–K compositions of all the compiled earth material samples. b) Triangular projection of the U–Th–K compositions of the muds and sands of recent worldwide deep-sea turbidites from McLennan et al. (1990).

In Fig. 3a we have projected the compiled RHPE concentrations in a triangular U–Th–K diagram. Most of the values fall close to the ratios Th/U = 4 and K/U = $1\times10^4$, though there is a considerable data dispersion. A similar distribution is observed within some lithological groups or specific lithotypes indicating the calculation of RHP values based on knowing the concentration of a unique radioelement must be taken with caution. An illustrative case of this is shown in Fig. 3b where the distribution of sand and mud fractions of worldwide deep-sea turbidites from McLennan et al. (1990) are plotted. Note that for both lithologies the ratio of Th/U vary around 8/1 and 5/4 and the ratio $K\times10^4$/U vary around 3/1 and 1/4.

Focusing on the main lithological groups and taking as a reference the RHP ranges defined by the 10th and the 90th percentiles, we observe that igneous rocks are those that present the largest variation of RHP (from 0.1 to 3.8 μW $m^{-3}$), followed by the metamorphic rocks (from 0.21 to 3.2 μW $m^{-3}$) and the sedimentary rocks (from 0.32 to 1.80 μW $m^{-3}$). The median RHP of sedimentary rocks (1 μW $m^{-3}$) is significantly lower than that of the igneous and metamorphic rocks (around 1.2 μW $m^{-3}$ in both cases) (Table 2). However, after considering the density (compaction) correction for sedimentary rocks, which amounts a factor between 1.14 and 1.20, we obtain similar RHP values for sedimentary, igneous and metamorphic rocks. In what follows we analize the variations in RHP obtained in each main lithological group and compare the obtained values with previous compilation studies.

### 4.1. Igneous rocks

Within this main group we distinguish between plutonic and volcanic rocks and their subgroups. Plutonic rocks exhibit a higher RHP variation and a higher RHP median and mean values than volcanic rocks. In the case of plutonic rocks a gradual RHP increasing can be observed from ultramafic rocks, to gabbroids, dioritoids, tonalites, granodiorites and granites. It is worth noting that a considerable overlapping in RHP values exists between these lithological groups. The ultramafic rocks compilation includes basically peridotites and, to a lesser extent, pyroxenites and hornblendites. Focusing on peridotites, the highest RHP values tend to be in enriched peridotites and the lowest ones in the depleted varieties. Based on the average composition of enriched peridotites proposed by McDonough (1990) and Rudnick et al. (1998) a RHP around 0.03 μW $m^{-3}$ is obtained, whereas depleted peridotites yield values around 0.002 μW $m^{-3}$. These estimates indicate that the compilation presented in our study is governed by enriched varieties provided that the median value is 0.111 μW $m^{-3}$. Pyroxenites and peridotites present similar RHP variabilities. The RHP for hornblendites tends to be higher than for pyroxenites and peridotites. The RHP ranges of gabbroids and dioritoids have an important overlapping. The RHP values of the compiled anorthosites fall inside the RHP range defined by the gabbroids. The mean and median RHP values of gabbroids are fairly close to those previously proposed by Haack (1983), while in the case of dioritoids they are significantly lower. Granitoids present a higher RHP variability than gabbroids and dioritoids. The median RHP of tonalites that we have obtained is very similar to that calculated from the average composition of tonalites proposed by Wedepohl (1995). In the case of granodiorites it must be highlighted that its RHP variability define a field that overlaps the RHP ranges of tonalites and granites. Granites present a high RHP variability, which is also observed in the different granite types (e.g. leucogranites, S-type). For instance, the compilation of 148 compositions of A-type granites (anorogenic) carried out by Whalen et al. (1987), results in a range of RHP between 0.52 and 12.7 μW $m^{-3}$ but a mean RHP value of 3.3 μW $m^{-3}$ with a standard deviation of 1.6 μW $m^{-3}$. This high RHP variability is related to the granite petrogenesis. Systematic variations in RHPE concentrations between and within granitic bodies (and in general inside granitoid complexes) are related to both, initial magma composition and magmatic differentiation processes.

Volcanic rocks show a clear gradual increase in RHP from basalt, to andesite and to rhyolite. Basalts show similar RHP values than gabbroids. The basalt lithological group includes alkaline, calc-alkaline and tholeiitic varieties the first variety being the most heat producing and the latest the lower. Based on the average compositions of basalts proposed by Sun and McDonough (1989), Taylor and McLennan (1985), Jochum et al. (1983), Sun (1980), Klein (2003) and Kelemen et al. (2003) we calculated a mean RHP of 0.7 μW $m^{-3}$, 0.3 μW $m^{-3}$ and 0.05 μW $m^{-3}$ for the alkalic basalts, the calc-alkaline basalts, and the tholeiitic basalts, respectively. The median RHP of andesites and rhyolites are similar to the average values proposed by Haack (1983). Interestingly, andesites show similar 50th, 25th and 75th RHP percentiles than dioritoids, whereas representative RHP values of rhyolites and granites are also similar. These similar RHP values are a consequence of the geochemical affinities between volcanic and plutonic rocks. Considering the petrologic fields that define the TAS diagram (the classification scheme of the volcanic rocks), we expect that: (a) dacites have a range of RHP values intermediate between the rhyolites and the andesites, and (b) basaltic andesites have a range of RHP values intermediate between andesites and basalts.

Taking as a reference the average major element chemical composition of the igneous rocks proposed by Le Maitre (1976), a correlation between RHP and $SiO_2$ content is obtained. According to Wollenberg and Smith (1987) this relationship is not maintained by including alkaline igneous rocks (with these names we include syenitoids, foid syenitoids, foid dioritoids, foid gabbroids, foidolites, trachitoids, phonolitoids, tephritoids and foiditoids). These alkaline rocks make up a small percentage of all the igneous rocks. According to Wollenberg and Smith (1987) alkalic intermediate varieties show a high RHP variability with values ranging between more than 10 μW $m^{-3}$ and less than 1 μW $m^{-3}$, whereas alkalic basic varieties show less variability, with values usually below 4 μW $m^{-3}$.

### 4.2. Sedimentary rocks

In the case of the sedimentary rocks, density plays an important role due to compaction. The values of RHP that we propose are based on a density of 2400 kg $m^{-3}$ and rarely exceed 3 μW $m^{-3}$. Only some rare varieties such as organic rich sediments, K-rich evaporites and phosphate-rich sediments have systematic higher RHP values, exceeding in some cases the value of 5 μW $m^{-3}$. When comparing with previous studies referred to sedimentary rocks, we have previously weighted the reported values to a common density of 2400 kg $m^{-3}$.

In detritic sedimentary rocks we observe that RHP is correlated with the grain size such that the higher the grain size the lower the RHP. As illustrated in Table 2, RHP decreases from mudrocks (median RHP = 1.442 μW $m^{-3}$), to wackes (median RHP = 0.993 μW $m^{-3}$) and to sandstones (median RHP = 0.819 μW $m^{-3}$). Bearing in mind the petrogenesis of coarse grained detritic materials we expect the RHP of conglomerates to be similar to sandstones. By mudrocks we have included all kinds of fine-grained detritic sedimentary rocks disregarding whether they are consolidated or unconsolidated (mudrock, shale, argillite, siltstone, mud, lutite, clay, silt). The different average shale compositions proposed by previous authors (e.g., Clark, 1966; Haack, 1983; Taylor and McLennan, 1985; Rybach, 1986; and Condie, 1993) give a RHP between 1.07 and 1.86 μW $m^{-3}$ with an average value of ~1.7 μW $m^{-3}$. This value is above the mean and median RHP values reported in our study for mudrocks (1.392 and 1.442 μW $m^{-3}$, respectively). The discrepancy is probably related to the different rock types included under the denomination of mudrocks used in our study and shales/sandstones used in previous studies. Within the sandstone group, arkoses generally present the highest values (around 1.25 μW $m^{-3}$) and quartz arenites the lowest ones (around 0.5 μW $m^{-3}$).

In general, carbonate-rich sedimentary rocks show a RHP lower than detritic sedimentary rocks, with values ranging between 0.1 μW $m^{-3}$ and 1 μW $m^{-3}$ and with no significant differences between limestones and dolostones. The obtained mean RHP of carbonate-rich sedimentary rocks

(0.447 μW m$^{-3}$ and median value of 0.416 μW m$^{-3}$) is slightly lower than that calculated from the average RHPE composition proposed by Wollenberg and Smith (1987), which results in a value of 0.65 μW m$^{-3}$. The average RHPE for limestones and dolostones proposed by Rybach (1986) yield values of 0.57 and 0.34 μW m$^{-3}$, respectively; whereas the average RHPE for limestones proposed by Haack (1983) results in 0.75 μW m$^{-3}$. In general evaporites (excluding KCl-rich varieties) show RHP values below 0.5 μW m$^{-3}$ as well as other minority sedimentary rock such as non-detritic siliceous and Fe–Mn-rich varieties.

### *4.3. Metamorphic rocks*

Metamorphic rocks show a great chemical diversity and consequently, a high RHP variability. Chemical composition of metamorphic rocks depends largely on the protolith from they belong, which may include igneous and sedimentary rocks. Furthermore, metamorphic fluid flow and partial melting processess can remobilize chemical elements (Ague, 1991; Chamberlain and Sonder, 1990) and therefore metamorphic rocks derived from a single protolith may show a wide range of RHP.

The low-to-medium grade metamorphic group yields a median and mean RHP of 1.5 μW m$^{-3}$. The corresponding cumulative plot (not shown) indicates that values between 0.1 and 3 μW m$^{-3}$ have the same probability to occur. Importantly, there are a large number of low grade metamorphic rocks, especially gneisses, which we were unable to classify as either metaigneous or metasedimentary. Metaigneous rocks generally show lower average values but higher variability than metasedimentary rocks. The mean and median RHP values of low grade metaigneous rocks differ significantly and are lower than the corresponding values for igneous rocks. The cumulative plot clearly shows two populations. The first one is defined by metamafic rocks (p. e. amfibolites, green schists) generally with values lower than 1 μW m$^{-3}$ and often below 0.1 μW m$^{-3}$, and the second one by metafelsic rocks (metagranitoids) generally with values higher than 1 μW m$^{-3}$ and often above 3 μW m$^{-3}$. The low grade metasedimentary rocks mainly include pelites, psamites, slates, phyllites, schists, marbles, metasandstones, metagreywackes and quartzites and general show higher RHP values than sedimentary rocks. These differences may be related to the increase of density due to compaction and the loss of sedimentary components depleted in RHPE (e.g. carbonates) during metamorphism. The range of RHP values for pelites is comparable to that for mudrocks, whereas the range of RHP values for psamites is similar to that for sandstones.

The group of high-grade metamorphic rocks includes those varieties related to partial melting processess or that have been subjected to *P*–*T* conditions over the granitic solidi. This lithological group is composed basically by migmatites, diatexites, metatexites, restites, mesosomes, melanosomes, granulites and eclogites. The RHP of this group is commonly lower than the low-to-medium grade metamorphic lithological group.

Granulites yield a very variable RHP with values corresponding to the 25th, 50th and 75th percentiles of 0.352, 0.753 and 1.613 μW m$^{-3}$, respectively (Table 2). These values are largely overestimated due to the bias introduced by sampling, which mainly correspond to outcropping felsic granulites. Intermediate and mafic granulites, which show a lower RHP than felsic granulites, are commonly sampled by xenoliths and then are scarcer. Based on the studies of Rudnick and Presper (1990), Rudnick (1992), Rudnick and Fountain (1995) and Rudnick and Gao (2003) we recommend an average RHP of 0.15 μW m$^{-3}$ for mafic granulites, 0.35 μW m$^{-3}$ for intermediate granulites and 0.85 μW m$^{-3}$ for felsic granulites. Eclogites show lower variability than granulites with a mean and a median value around 0.25 μW m$^{-3}$, similar to basalts.

## 5. Discussion

Establishing the RHP distribution in the lithosphere requires the previous knowledge of lateral and vertical variations of its composition. Our results show that each representative lithotype has a wide range of possible RHP values, which apart from nomenclature aspects, is mainly related to the different petrogenesis that a given lithotype can have. Basic petrologic premises concerning the processes that intervene in the formation and evolution of the continental crust and hence in the distribution of RHPE within the lithosphere are as follows:

1) From a petrological, geochemical and geophysical point of view, mantle rocks differ clearly from crustal rocks, both oceanic and continental. The upper mantle is believed to be mainly composed of peridotites (e.g. Anderson, 1983; Anderson, 1989; McDonough, 1990; McDonough and Sun, 1995). Eclogites, hornblendites and pyroxenites are other lithologies subordinated to peridotites in the lithospheric mantle. Only some ultramafic rocks derived by fractional crystallization of basaltic magmas can be misrelated to typical lithologies of the upper mantle.
2) Extensive laboratory experiments show that primary magmas derived from partial melting of the mantle have a basaltic composition (e.g. Wilson, 1989 and references therein) and crystallize into basalts or gabbros depending on the texture and the context where they crystallize. Knowing the RHP of basalts becomes essential to understand the crustal RHP distribution since the geochemical models for the formation and evolution of the continental crust assume an initial basaltic composition of oceanic origin (Wedepohl, 1995; Taylor and McLennnan, 1995; Rudnick, 1995).
3) Most of granitoid magmas that are emplaced into the upper continental crust or erupted from silicic volcanoes were generated during high-*T* (upper amphibolite- to granulite-facies metamorphism) crustal melting (Clemens, 1990; Clemens and Watkins, 2001). The link between crustal melting and granitoid (tonalitic, granodioritic and granitic) magma genesis is firmly established on the basis of the stable and radiogenic isotope characteristics of granitoid rocks, field studies of high-grade terranes and high *P*–*T* experimental studies. Metatonalites produced by dehydration melting of metabasalts are suitable parent rocks for the generation of granodioritic and granitic melts (Johannes and Holtz, 1995 and references therein). Though, granites and granodiorites may also be formed by partial melting of other crustal rocks such as pelites or greywackes.
4) Each protolith can yield different types of partial melting and melt segregation depending on the prevailing *P*–*T* conditions and other physico-chemical variables (Maaloe, 1982; Sawyer, 1996; Vigneresse et al., 1996). The partial melt product can react and equilibrate continuously with the crystalline residue until its segregation or can be continuously removed from the system as soon as it is formed, changing the bulk composition of the system. Melt produced by partial melting can segregate alone or with a fraction of the residuum. These different types of magma genesis contribute to the RHP variability observed in the magmatic lithological groups.
5) After primary magmas segregate from their source region they may undergo a variety of complex fractionation, mixing and contamination processes during transport and subsequent storage in higher crustal levels (e.g. Wilson, 1989). These processes are of fundamental importance in producing the RHP variabilty found in the different magmatic rocks.
6) Apart from the partial melting processes that affect high-grade metamorphic rocks, large-scale fluid flow processes can remobilize RHPE during a wide range of non-anatectic metamorphic conditions (Chamberlain and Sonder, 1990; Ague, 1991).
7) Chemical and physical weathering, mineral sorting during transport and diagenesis fractionate RHPE during sedimentary material formation (McLennan, 2001; Rudnick and Gao, 2003). It is important to note that U and K have a high potential of being fractionated during sedimentary processes because they have high solubilities in natural waters (Taylor and McLennan, 1985).

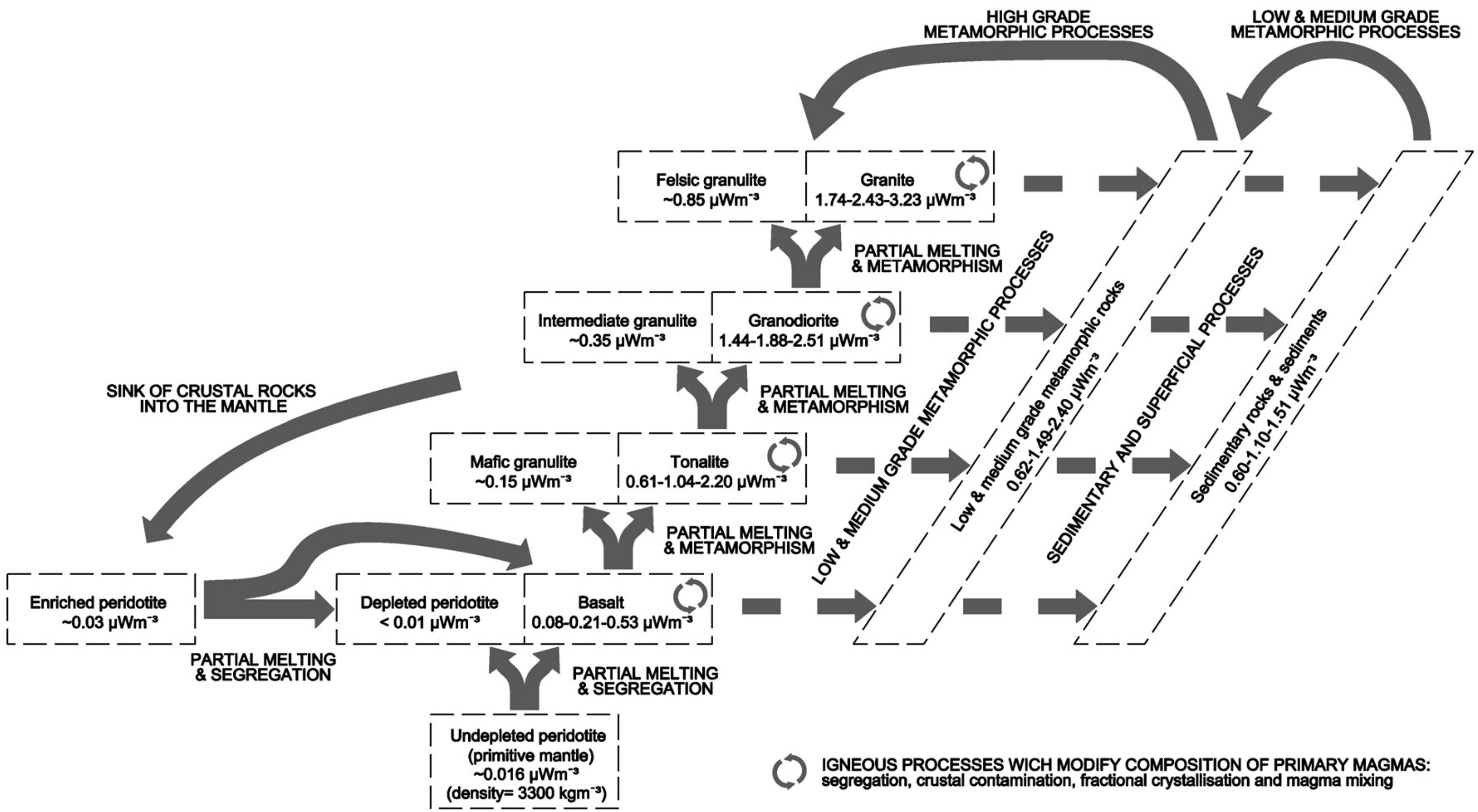


**Fig. 4.** Schematic diagram of the differentiation processes undergone by lithospheric rocks. At a given step the diagram shows the segregated product of a partial melting process (arrows pointing to upper-right) and the corresponding restite (arrows pointing to upper-left). The segregated product can undergo low or medium metamorphism and eventually to become sedimentary rocks (arrows pointing to right). Sedimentary rocks can in turn be converted to low–medium metamorphic rocks or to undergo high-grade metamorphism with partial melting, as indicated by the uppermost arrows. Subduction and melting of crustal rocks into the deep mantle produces enriched peridotites which can be partly melted being incorporated again into the cycle.

*5.1. Proposed radiogenic heat production values*

Bearing in mind the various differentiation processes above mentioned, we have constructed a schematic diagram (Fig. 4) that summarizes the representative RHP values to be adscribed to each rock-type and the criteria for their correct use particularly in thermal modeling purposes. In this diagram we have included the main lithotypes described in Table 2 and the RHP values corresponding to the 25th, 50th and 75th percentiles.

According to Taylor and McLennnan (1995) the continental crust grew through geological time and begins its evolution with a thick basaltic (oceanic) crust derived from peridotite partial melting (undepleted mantle), which is gradually replaced by a more radiogenic (granitic) component. The evolution of the continental crust into a more rich radiogenic composition is due to the sink of dense and RHP depleted crustal material into the upper mantle (Johannes and Holtz, 1995). The intrusion of melts derived from partial melting of subducted crust also enriches the continental crust in RHP. In the zones in which the crustal material comes back to the mantle, such as the subduction or delamination zones, mantle rocks become enriched in RHPE. This recycling process that affects the crust and the upper mantle is denoted by the lower panels of our scheme in Fig. 4. Therefore, appropriate RHP values for mantle peridotites range from $<0.01\ \mu W\ m^{-3}$ for depleted mantle, to $\sim 0.016\ \mu W\ m^{-3}$ for primitive-undepleted mantle and to $0.03\ \mu W\ m^{-3}$ for enriched mantle. The basalts resulting from partial melting and segregation of mantle rocks have a median RHP value of $0.21 \pm [0.08-0.53]\ \mu W\ m^{-3}$, where the numbers in brackets denote the values corresponding to 25th and 75th percentiles, respectively.

The first intracrustal magmatic episode is the generation of tonalitic melts by partial melting of metabasalts. The resultant rock types generated through separation of melts and solids are respectively tonalities, with a median RHP of $1.04 \pm [0.61-2.20]\ \mu W\ m^{-3}$, and mafic granulites with an average RHP of $0.15\ \mu W\ m^{-3}$. Tonalites and sedimentary rocks such as mudstones and wackes are suitable parent rocks for the generation of granitic partial melts. This second intracrustal magmatic episode results in granodiorites with a median RHP of $1.88 \pm [1.44-2.51]\ \mu W\ m^{-3}$ and granites with a median RHP of $2.43 \pm [1.74-3.23]\ \mu W\ m^{-3}$, both lithotypes generally emplaced in the upper crust. The crustal residue composed of intermediate ($\sim 0.35\ \mu W\ m^{-3}$) to felsic ($\sim 0.85\ \mu W\ m^{-3}$) granulites is emplaced at the middle–lower crust. The analyses of surface heat flow and heat production data sets show that in most continental regions the upper crustal levels are significantly enriched in RHP relative to the deeper crustal levels (see Table 2 and Fig. 4). The evolution of the continental crust into a granodioritic upper crust and a lower crust dominated by intermediate and mafic granulites (Johannes and Holtz, 1995) is a consequence of the intracrustal magmatic episodes described previously. Granulite-facies metamorphism in the lower crust cause pervasive depletion of U, Th and K due to the breakdown of accessory phases, partial melting and removal of melts and fluids at higher crustal levels (Rudnick et al., 1998). The existence of mantle-derived rocks (gabbros) underplated in the lower crust also could favor the lower concentration of RHP in the deep crust with respect to the shallower crustal levels where metasedimentary and more evolved metaigneous rocks are present. However, the presence of a mafic layer at the base of the continental crust is not a universal feature (Wedepohl, 1995; Villaseca et al., 1999), since delamination and detachment of the lower crust are likely to occur (Bird, 1979; Rudnick, 1995; Schott and Schmeling, 1998).

Igneous rocks can undergo metamorphic processes of different grade and/or be converted in sedimentary rocks. In turn, sedimentary rocks can be transformed to low–medium grade metamorphic rocks. In average, low–medium grade metamorphic rocks show a median RHP of $1.49 \pm [0.62-2.40]\ \mu W\ m^{-3}$, whereas sedimentary rocks show a median value of $1.10 \pm [0.60-1.51]\ \mu W m^{-3}$. Note that none of the values reported in Fig. 4 are reduced to a reference density.

*5.2. Thermal modeling*

To analyze how departures from the median values (25th and 75th percentiles) can affect the crustal geotherms we have performed a simple calculation based on a 1D steady-state thermal approach. To this end we have considered a continental crustal section derived by Wedepohl (1995) who proposes a crustal composition based on the proportions of upper crust to felsic lower crust to mafic lower crust of 1:0.6:0.4. These proportions were derived from a 3000 km long refraction seismic profile through western Europe (EGT-European GeoTraverse) comprising 60% old shield and 40% younger fold belt with an average crustal thickness of about 40 km. We have chosen this section because of the precise description of the average crustal composition, which is based on worldwide mapping, petrologic studies and chemical balances. Therefore, the crust consists of three layers, namely sediments, upper crust and lower crust whose composition and thickness are summarized in Table 3. For the lithotypes composing the

**Table 3**
Radiogenic heat production (RHP) and radiogenic heat generation (RHG) obtained for a standard crustal column according to Wedepohl (1995).

| Composition (%) and depth range | Lithology | Thickness km | RHP $\mu W\ m^{-3}$ | RHG $mW\ m^{-2}$ |
|---|---|---|---|---|
| Sediments (0–3 km) | | | | |
| 44.2 | Mudrocks | 3.00 | 1.44 | 1.90 |
| 20.9 | Sandstone-wackes | 3.00 | 0.90 | 0.56 |
| 20.3 | Mafic volcanics | 3.00 | 0.21 | 0.13 |
| 14.6 | Carbonates | 3.00 | 0.42 | 0.18 |
| Total | | 3.00 | 0.93 | 2.78 |
| Upper crust (3–21 km) | | | | |
| 28.9 | Granites | 18.00 | 2.43 ± [1.74–3.23] | 12.63 ± [9.05–16.81] |
| 23.1 | Granodiorites | 18.00 | 1.88 ± [1.44–2.51] | 7.82 ± [5.98–10.45] |
| 5.8 | Tonalites | 18.00 | 1.04 ± [0.61–2.20] | 1.08 ± [0.64–2.28] |
| 7.2 | Gabbros | 18.00 | 0.35 ± [0.18–0.72] | 0.45 ± [0.23–0.94] |
| 35.0 | Low–medium grade metamorphics | 18.00 | 1.49 ± [0.62–2.39] | 9.39 ± [3.90–15.06] |
| Total | | 18.00 | 1.74 ± [1.10–2.53] | 31.37 ± [19.80–45.54] |
| Lower crust (21–40 km) | | | | |
| 61.5 | Felsic-intermediate granulites | 19.00 | 0.50 | 5.84 |
| 38.5 | Mafic granulites | 19.00 | 0.15 | 1.10 |
| Total | | 19.00 | 0.37 | 6.94 |
| Total crust (0–40 km) | | 40.00 | 1.03 ± [0.74–1.38] | 41.09 ± [29.52–55.26] |

RHP and RHG denotes radiogenic heat production and radiogenic heat generation calculated for the 50th percentile (median value), respectively. Values between brackets correspond to 25th and 75th percentiles.

sedimentary layer and the lower crust, we have considered a RHP according to the median values reported in Table 2 and Fig. 4. However, for the upper crust we have considered the median RHP values and the range of variation corresponding to the 25th and 75th percentiles. The equivalent RHP and the total amount of heat generated in the three layers are displayed in Table 3. According to this, the bulk crustal RHP can vary from 0.74 μW m$^{-3}$ to 1.38 μW m$^{-3}$ being the most probable value 1.03 μW m$^{-3}$. These values are well inside the range reported for two extreme cases of high and low average crustal RHP as are the Trans-Hudson Orogen with a value of 0.55 μW m$^{-3}$ (Perry et al., 2006) and the Australian Proterozoic crust with a value greater than 2 μW m$^{-3}$ (Hand et al., 1999). Our values are about 0.15 μW m$^{-3}$ higher than those reported for different continental crust compositional models (Rudnick and Gao, 2003 and references therein), which give a range of mean crustal RHP values from 1.29 to 0.57 μW m$^{-3}$. The reason for this shifting is the relatively felsic composition of the crustal section considered by Wedepohl (1995). Radiogenic heat production in the lithospheric mantle has been set to 0.016 μW m$^{-3}$, which corresponds to a primitive (undepleted) mantle composition (Fig. 4).

1D thermal calculations have been firstly performed considering as boundary conditions a surface temperature of 0 °C and a surface heat flow of 65 mW m$^{-2}$ and using the analytical solution $T(z) = T_0 + q_0 z/\lambda - Az^2/2\lambda$ where $T_0$ and $q_0$ are the temperature and heat flow at the Earth's surface, respectively, $A$ is the (constant) radiogenic heat production, and λ is the thermal conductivity (e.g. Turcotte and Schubert, 2002). Bulk thermal conductivity of sediments, upper crust, lower crust and lithospheric mantle are considered constant with values of 2.5, 3.0, 2.1 and 3.4 Wm$^{-1}$ K$^{-1}$, respectively. Calculated temperatures (Table 4) show large differences related to the considered variations in the upper crust RHP, ranging from 313 °C to 390 °C at the base of the upper crust and from 433 °C to 743 °C at the Moho. Similarly, the obtained heat flow at the base of the crust and the total lithospheric thickness, here defined as the depth of the 1350 °C isotherm, vary from 10 to 35 mW m$^{-2}$ and from >400 km to 99 km, respectively (Table 4). These extreme geotherms show $T_{\text{Moho}}$, mantle heat flow and lithospheric thickness values that cover most of the range of characteristic values for cratonic and warm/young continental lithospheres (e.g., Jaupart and Mareschal, 2007).

Using surface heat flow as boundary condition tends to transmit downwards the uncertainties in RHP. However, fixing the lithospheric thickness instead of surface heat flow reduces considerably the variations of the calculated geotherms. We have used a FEM based numerical solution that allows us to calculate the temperature distribution in a multilayer media with the same prescribed thicknesses of the crustal layers and thermal properties as in the analytical case but fixing the temperatures at the surface and base of the lithosphere and the total lithospheric thickness. Then, using $T_0 = 0$ °C, $T_B = 1350$ °C and a lithospheric thickness of 150 km as boundary conditions results in Moho temperatures and surface heat flow values ranging from 571 °C to 642 °C and from 54 mW m$^{-2}$ to 78 mW m$^{-2}$ for 25th and 75th RHP percentiles, respectively (Table 4). The mantle heat flow remains almost constant around 22 mW m$^{-2}$. Whatever the boundary conditions are used, the resulting uncertainties in the calculated geotherms and the surface heat flow or lithospheric thickness are large. It is worth noting that variations in the calculated geotherms do not depend only on radiogenic heat production but also on thermal conductivity and lithosphere geometry. More sophisticated models based on Monte-Carlo inversion techniques help to better define the continental geotherms by considering random variations in RHP, thermal conductivity and thickness of the lithosphere layers (e.g., Jokinen and Kukkonen, 1999a, b). The use of surface heat flow or lithospheric thickness as boundary conditions will depend on the data available in each case. Surface heat flow is commonly determined with 10–15% accuracy (e.g. Powell et al., 1988) resulting in a wide range of possible geotherms. This range can be additionally constrained by integrated modeling techniques that combine data from petrology, mineral physics, and geophysical observables within a self-consistent framework. The final result is a lithosphere/sublithospheric model that simultaneously fits all geophysical observables (gravity, geoid, elevation, surface heat flow, seismic velocities and composition) reducing the uncertainties associated with thermal modeling alone (e.g., Afonso et al., 2008; Fullea et al., 2009).

### 5.3. Radiogenic heat production versus age

The heterogeneity of the continental crust records the complex superposition of tectonic processes associated with the growth and differentiation of the continental lithosphere through age. Reviews on heat flow data of stable continent regions recognize a decrease in the global continental heat flow with age (Nyblade and Pollack, 1993). Differences in the concentration of RHPE of crustal units and its thickness are the two possible causes of the temporal pattern of the RHP in the crust (Nyblade and Pollack 1993; Rudnick and Fountain, 1995; McLennan and Taylor, 1996; Rudnick et al., 1998; Nyblade, 1998). Focusing on the RHP changes and according to Jaupart and Mareschal (2003), estimates of bulk continental RHP from heat flow data shows that there is a clear decrease with age.

Unfortunately, our database does not contain enough details to segregate the rock samples by age intervals. To explore the age dependence of RHP we have calculated the RHP values of published average chemical compositions representative of crustal lithotypes grouped by age. Then, we have used the worldwide chemical compositions of representative upper crust lithotypes by Condie (1993), the average chemical compositions representative of Central–East China by Gao et al. (1998), and the worldwide average chemical compositions of different types of granulites by Rudnick and Fountain (1995). Fig. 5 compares the calculated RHP values to the age of the different

**Table 4**
Calculated temperatures and heat flow values from models considering constant surface heat flow and constant lithospheric thickness as boundary conditions. Radiogenic heat production (RHP) in the upper crust varies according to 25th, 50th and 75th percentiles.

| | Depth (km) | Constant surface heat flow | | | | | | Constant lithospheric thickness | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | Avg RHP | | Min RHP | | Max RHP | | Avg RHP | | Min RHP | | Max RHP | |
| | | Temper. (°C) | Heat flow (mW m$^{-2}$) | Temper. (°C) | Heat flow (mW m$^{-2}$) | Temper. (°C) | Heat flow (mW m$^{-2}$) | Temper. (°C) | Heat flow (mW m$^{-2}$) | Temper. (°C) | Heat flow (mW m$^{-2}$) | Temper. (°C) | Heat flow (mW m$^{-2}$) |
| Surface | 0 | 0 | 65 | 0 | 65 | 0 | 65 | 0 | 65 | 0 | 54 | 0 | 78 |
| Base sedim. | 3 | 76 | 62 | 76 | 62 | 76 | 62 | 76 | 62 | 64 | 52 | 92 | 75 |
| Base U. crust | 21 | 356 | 31 | 390 | 42 | 313 | 17 | 356 | 31 | 314 | 32 | 406 | 30 |
| Base L. crust | 40 | 603 | 24 | 743 | 35 | 433 | 10 | 603 | 24 | 571 | 25 | 642 | 23 |
| Lith. thickness (km) and basal heat flow (mW m$^{-2}$) | | 150/22 | | 99/33 | | 447/8 | | 150/22 | | 150/23 | | 150/21 | |

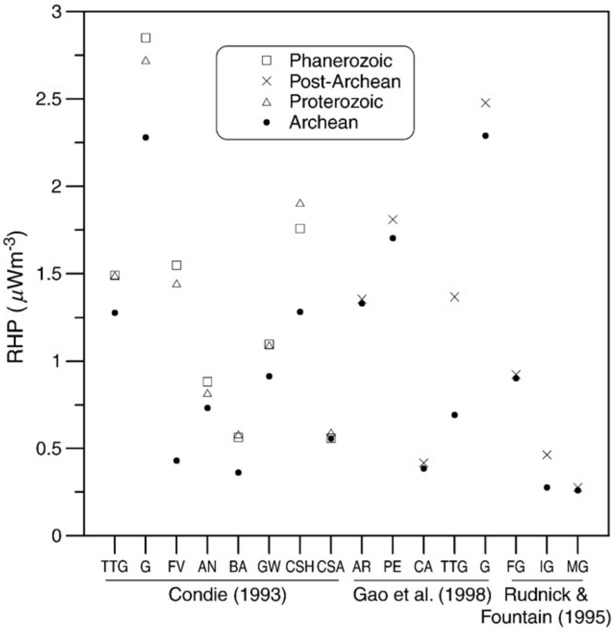


**Fig. 5.** RHP of different representative lithotypes of various ages. Tonalite–trondhjemite–granodiorite (TTG), granite (G), felsic volcanic rocks (FV), andesites (AN), basalts (BA), graywackes (GW), cratonic shales (CSH), cratonic sandstones (CSA) from Condie (1993); arenaceous rocks (AR), pelite (PE), carbonate (CA), Tonalite–trondhjemite–granodiorite (TTG) and granite (G) from central East China of Gao et al. (1998); and felsic granulites (FG), intermediate granulites (IG) and mafic granulites (MG) from worldwide compilation of Rudnick and Fountain (1995).

representative lithotypes. Clearly, RHP values of Archean samples are always lower than Post-Archean thus indicating a decrease RHP with age. RHP values of Proterozoic and Phanerozoic samples are much closer and therefore, this RHP–age relationship is not so clear (Fig. 5). Variations of RHP with age may amount to 20–50% or even a factor 3 in the case of felsic volcanic rocks. Only the samples corresponding to cratonic sandstones by Condie (1993), arenaceous and carbonate rocks by Gao et al. (1998) and felsic and mafic granulites by Rudnick and Fountain (1995) this RHP–age relationship is not so evident.

## 6. Concluding remarks

Based on a worldwide compilation of radiogenic heat production data and isotopic analyses we propose typical values of RHP for common lithologies and study its variability with the aim to be useful in lithospheric thermal modeling purposes. The main conclusions that are derived from the presented work can be summarized as follows:

i) Continental crust rocks show a high variability in RHP due to their petrogenesis and to the complex differentiation and redistribution processes affecting the whole lithosphere. The trace element character of U and Th and their association with accessory minerals make difficult to establish a direct relationship between RHP and lithology. All these impede to find a simple depth-dependent function for RHP within the continental crust.

ii) For modeling purposes it is essential first to identify the lithologies to be modeled, which must be referred to the hierarchical petrologic-based classification summarized in this work. This classification includes 24 representative crust and upper mantle lithologies grouped into three main lithological groups encompassing igneous, metamorphic and sedimentary rocks.

iii) To minimize the influence of outliers and departures from a normal statistical distribution within each lithological group we propose to use, instead of the mean value and its standard deviation, the median value (50th percentile) and the range of values defined by the 25th and 75th percentiles to characterize the RHP of each lithotype and its variability.

iv) From the compiled data we observe that the largest variation in bulk RHP corresponds to igneous rocks (from 0.1 to 3.8 μW m$^{-3}$), followed by metamorphic rocks (from 0.2 to 3.2 μW m$^{-3}$) and sedimentary rocks (from 0.3 to 1.8 μW m$^{-3}$) the later showing an important bias due to the lower bulk density. Igneous rocks and their metamorphic and sedimentary derivates show a trend to increase its RHPE concentrations as they melt and segregate from the depleted restite.

v) The RHP variability within the upper crust of a typical crustal section can produce noticeable temperature and heat flow variations in the crust–mantle boundary amounting up to 300 °C and 25 mW m$^{-2}$, respectively. These variations are sensitive to the applied boundary conditions and reduce considerably when other geophysical/petrological observables are incorporated to thermal models.

vi) RHP tends to decrease with age and these variations may amount 20–50% when comparing Archean to Post-Archean rocks. Most of the sedimentary rocks as well as felsic and mafic granulites show negligible variations whereas maximum variations by a factor of 3 are found in felsic volcanic rocks.

vii) Implementing radiogenic heat production in modeling the thermal structure of the lithosphere requires (a) to identify the main lithologies and its proportion (if possible) composing the different layers; (b) to consider the range of variability of RHP values and the age of the crustal domains; and (c) to integrate as many as possible additional constraints that allow us to better define the geometry and boundary conditions of the lithosphere section. In this sense, the use of integrated lithospheric models and/or inversion methods that account for the variability of the different physical parameters intervening in the model is highly recommended. RHP measurements in rock samples derived from the modeling region are essential.

## Acknowledgements

This research was partly financed by Projects TopoMed-Spain (CGL2008-03474-E/BTE), ATIZA (CGL2009-09662/BTE) and the Consolider-Ingenio 2010 Topo-Iberia (CSD2006-00041). We thank the editors Mike Sandiford and Fabrizio Storti, and Ilmo Kukkonen and an anonymous reviewer for their very constructive comments on the first version of the manuscript.

## Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.tecto.2010.05.003.

## References

Afonso, J.C., Fernàndez, M., Ranalli, G., Griffin, W.L., Connolly, J.A.D., 2008. Integrated geophysical–petrological modelling of the lithosphere and sublithospheric upper mantle: methodology and applications. Geochemistry Geophysics Geosystems 9, Q05008. doi:10.1029/2007GC001834.
Ague, J.J., 1991. Evidence for major mass transfer and volume strain during regional metamorphism of pelites. Geology 19, 855–858.
Anderson, D.L., 1983. Chemical composition of the mantle. Journal of Geophysical Research B 88, 41–52.
Anderson, D.L., 1989. Theory of the Earth. Blackwell, Boston, p. 366.
Artemieva, I.M., Mooney, W.D., 2001. Thermal thickness of Precambrian lithosphere: a global study. Journal of Geophysical Research 106 (B8), 16,387–16,414.
Beardsmore, G.R., Cull, J.P., 2001. Crustal Heat Flow: A Guide to Measurement and Modelling. Cambridge University Press, Cambridge. pp. 324.
Birch, F., Roy, R.F., Decker, 1968. Heat flow and thermal history in New York and New England. In: Zen, E., White, W.S., Hadley, J.B., Thompson, J.B. (Eds.), Studies of Appalachian Geology. Northern and Maritime, Interscience, New York, pp. 437–451.
Bird, P., 1979. Continental delamination and the Colorado Plateau. Journal of Geophysical Research 84, 7561–7571.

Brady, R.J., Ducea, M.N., Kidder, S., Saleeby, J., 2006. The distribution of radiogenic heat production as a function of depth in the Sierra Nevada Batholith, California. Lithos 86, 229–244.

Chamberlain, C.P., Sonder, L.J., 1990. Heat-producing elements and the thermal and baric patterns of metamorphic belts. Science 250, 763–769.

Christensen, N.I., Mooney, W.D., 1995. Seismic velocity structure and composition of the compostion of the continental crust: a global review. Journal of Geophysical Research 100, 9761–9788.

Clark, S.P., 1966. Handbook of physical constants. Geological Society of America Memoir, 97. Boulder, p. 587.

Clemens, J.D., 1990. The granulite–granite connexion. In: Vielzeuf, D., Vidal, P. (Eds.), Granulites and Crustal Differentiation. Kluwer, Dordrecht, pp. 25–36.

Clemens, J.D., Watkins, J.M., 2001. The fluid regime of high-temperature metamorphism during granitoid magma genesis. Contributions to Mineralogy and Petrology 140, 600–606.

Condie, K.C., 1993. Chemical composition and evolution of the upper continental crust: contrasting results from surface samples and shales. Chemical Geology 104, 1–37.

Darnley, A., 1995. A Global Geochemical Database for Environmental and Resource Management. UNESCO Publishing, Ottawa. pp. 122.

Fountain, D.M., 1985. Is there a relationship between seismic velocity and heat production for for crustal rocks? Earth and Planetary Science Letters 79, 145–150.

Fountain, D.M., Furlong, K.P., Salisbury, M.H., 1987. A heat production model of a shield area and its implications for the heat flow-heat production relationship. Geophysical Research Letters 14, 283–286.

Fullea, J., Afonso, J.C., Conolly, J.A.D., Fernandez, M., Garcia-Castellanos, D., Zeyen, H., 2009. LitMod3D: an interactive 3-D software to model the thermal, compositional, density, seismological, and rheological structure of the lithosphere and sublithospheric mantle. Geochemistry Geophysics Geosystems 10, Q05019. doi:10.1029/2009GC002391.

Furlong, K.P., Chapman, D.S., 1987. Crustal heterogeneities and the thermal structure of the continental crust. Geophysical Research Letters 14, 314–317.

Gao, S., Luo, T.-C., Zhang, B.-R., Zhang, H.-F., Han, Y.-W., Hu, Y.-K., Zhao, Z.-D., 1998. Chemical composition of the continental crust as revealed by studies in East China. Geochimica et Cosmochimica Acta 62, 1959–1975.

Haack, U., 1983. On the content and vertical distribution of K, Th and U in the continental crust. Earth and Planetary Science Letters 62, 360–366.

Hand, M., Sandiford, M., Wyborn, L., 1999. Some thermal consequences of high heat production in the Australian Proterozoic. AGSO Research Newsletter 30.

Hofmann, A., 1988. Chemical differentiation of the Earth: the relationship between mantle, continental crust, and oceanic crust. Earth and Planetary Science Letters 90, 297–314.

IAEA, 2003. Guidelines for Radioelement Mapping Using Gamma Ray Spectrometry Data. TECDOC-1363. International Atomic Energy Agency, Vienna.

Jaupart, C., Mareschal, J.C., 2003. Constraints on crustal heat production from heat flow data. pp. 65–84. In: the Crust (Ed. R.L. Rudnick) Vol. 3 Treatise on geochemistry (Eds. H.D. Holland and K.K. Turekian). Elsevier-Pergamon, Oxford.

Jaupart, C., Mareschal, J.C., 2007. Heat flow and thermal structure of the lithosphere data. pp. 217–251. In: Crustal and lithosphere dynamics (Ed. A.D. Watts) Vol. 6 Treatise on geophysics (Ed. G. Schubert). Elsevier, Oxford.

Jochum, K.P., Hofmann, A.W., Ito, E., Seufert, H.M., White, W.M., 1983. K, U and Th in mid-ocean ridge basalt glasses and heat production, K/U and K/Rb in the mantle. Nature 306, 431–436.

Joeleht, A., Kukkonen, I., 1998. Thermal properties of granulite facies rocks in the Precambrian basement of Finland and Estonia. Tectonophysics 291, 195–203.

Jokinen, J., Kukkonen, I., 1999a. Random modelling of the lithospheric thermal regime: forward simulations applied in uncertainty analysis. Tectonophysics 306, 277–292.

Jokinen, J., Kukkonen, I., 1999b. Inverse simulation of the lithospheric thermal regimes using the Monte Carlo method. Tectonophysics 306, 293–310.

Johannes, W., Holtz, F., 1995. Petrogenesis and Experimental Petrology of Granitic Rocks. Springer, Berlin, p. 335.

Johnson, C.R., Olhoelf, G.R., 1982. Density of rocks and minerals. In: Charmichael, R.S. (Ed.), Handbook of Physical Properties of Rocks VIII. Press, CRC, pp. 1–39.

Kelemen, P.B., K. Hanghoj, K., Greene, A.R., 2003. One view of the geochemistry of subduction-related magmatic arcs with an emphasis on primitive andesite and lower crust. pp, 593–659. In the Crust (Ed. R.L. Rudnick) Vol. 3 Treatise on geochemistry (Eds. H.D. Holland and K.K. Turekian) Elsevier-Pergamon, Oxford.

Klein, E.M., 2003. Geochemistry of the igneous oceanic crust. pp 433–463. In the Crust (Ed. R.L. Rudnick) Vol. 3 Treatise on geochemistry (Eds. H.D. Holland and K.K. Turekian) Elsevier-Pergamon, Oxford.

Kukkonen, I., Lahtinen, R., 2001. Variation of radiogenic heat production rate in 2.8-1.8 Ga old rocks in the central Fennoscandian shield. Physics of the Earth and Planetary Interiors 126, 279–294.

Kukkonen, I., Peltoniemi, S., 1998. Relationships between thermal and other petrophysical properties of rocks in Finland. Physics and Chemistry of the Earth 23, 341–349.

Lachenbruch, A.H., 1968. Preliminary geothermal model of the Sierra Nevada. Journal of Geophysisical Research 73, 6977–6989.

Le Maitre, R.W., 1976. The chemical variability of some common igneous rocks. Journal of Petrology 17, 589–637.

Le Maitre, R.W., 2003. A Classification of Igneous Rocks and Glossary of Terms, 2nd Edition. Cambridge University Press, p. 224.

Lewis, T.J., Hyndman, R.D., Flück, P., 2003. Heat flow, heat generation, and crustal temperatures in the northern Canadian Cordillera: thermal control of tectonics. Journal of Geophysical Research 108 (B6), 2316. doi:10.1029/2002JB002090.

Maaloe, S., 1982. Geochemical aspects of permeability controlled partial melting and fractional crystallisation. Geochimica et Cosmochimica Acta 46, 43–57.

Marchalland, C.P., Fairbridge, R.W., 1999. Encyclopedia of Geochemistry. Kluwer Academic Press.

McDonough, W.F., 1990. Constraints on the composition of the continental lithospheric mantle. Earth and Planetary Science Letters 101, 1–18.

McDonough, W.F., Sun, S., 1995. The composition of the Earth. Chemical Geology 120, 223–253.

McLennan, S.M., Taylor, S.R., McCulloch, M.T., Maynard, J.B., 1990. Geochemical and Nd–Sr isotopic composition of deep-sea turbidites: crustal evolution and plate tectonic association. Geochimica et Cosmochica Acta 54, 2015–2050.

McLennan, S.M., Taylor, S.R., 1996. Heat flow and the chemical composition of the continental crust. Journal of the Geological Society 104, 369–377.

McLennan, S.M., 2001. Relationship between the trace element composition of sedimentary rocks and upper continental crust. Geochemistry Geophysics Geosystems 2 (article 2000GC000109).

Morgan, P., Sawka, W.N., Furlong, K.P., 1987. Introduction: background and implications of the linear heat flow–heat production relationship. Geophysical Research Letters 14, 248–251.

Nyblade, A.A., Pollack, H.N., 1993. Can differences in heat flow between east and southern Africa be easily interpreted?: implications for understanding regional variability in continental heat flow. Tectonophysics 219, 257–272.

Nyblade, A.A., 1998. Heat flow and the structure of Precambrian lithosphere. Lithos 48, 81–91.

Palme, H., O'Neill, H.S., 2003. Cosmochemical estimates of mantle composition, pp, 1–38 In the The mantle and core (Ed. R.W. Carlson) Vol 2. Treatise on geochemistry (Eds. H.D. Holland and K.K. Turekian), Elsevier-Pergamon, Oxford.

Perry, H.K.C., Jaupart, C., Mareschal, J.-C., Bienfait, G., 2006. Crustal heat production in the Superior Province, Canadian Shield, and in North America inferred from heat flow data. Journal of Geophysical Research 111, B04401. doi:10.1029/2005JB003893.

Plant, J.A., Simpson, P.R., Smith, B., Windley, B.F., 1999. Uranium ore deposits—products of the radioactive Earth. In: Burns, P.C., Finch, R. (Eds.), Uranium: Mineralogy, Geochemistry and Environment: Reviews in Mineralogy, 38, pp. 255–319.

Pollack, H.N., Hurter, S.J., Johnson, J.R., 1993. Heat flow from the Earth's interior analysis of the global data set. Reviews of Geophysics 31, 267–280.

Powell, W.G., Chapman, D.S., Balling, N., Beck, A.E., 1988. Continental heat-flow density. In: Haenel, R., Rybach, L., Stegena, L. (Eds.), Handbook of Terrestrial Heat-Flow Density Determination. Kluwer Ac. Publ, Dordrecht, pp. 167–222.

Roy, R.F., Blackwell, D.D., Birch, F., 1968. Heat generation of plutonic rocks and continental heat flow provinces. Earth and Planetary Science Letters 5, 1–12.

Rudnick, R.L., Presper, T., 1990. Geochemistry of intermediate to high-pressure granulites. In: Vielzeuf, D., Vidal, P. (Eds.), Granulites and Crustal Evolution. Kluwer, pp. 523–550.

Rudnick, R.L., 1992. Xenoliths — samples of the lower crust. In: Fountain, D.M., Arculus, R., Kay, R.W. (Eds.), Continental Lower Crust. Elsevier, Amsterdam, pp. 269–316.

Rudnick, R.L., 1995. Making continental crust. Nature 378, 571–578.

Rudnick, R.L., Fountain, D.M., 1995. Nature and composition of the continental crust: a lower crustal perspective. Reviews of Geophysics 33, 267–309.

Rudnick, R.L., McDonough, W., O'Conell, R.J., 1998. Thermal structure, thickness and composition of continental lithosphere. Chemical Geology 145, 395–411.

Rudnick, R.L., Gao, S., 2003. Composition of the continental crust, pp. 1–64. In the Crust (Ed. R.L. Rudnick) Vol. 3 Treatise on geochemistry (Eds. H.D. Holland and K.K. Turekian). Elsevier-Pergamon, Oxford.

Rybach, L., 1976. Radioactive heat production: a physical property determined by the chemistry of rocks. In: Sterns, R.G.J. (Ed.), The Physics and Chemistry of Minerals and Rocks. Wiley and Sons, pp. 309–318.

Rybach, L., Buntebarth, G., 1982. Relationship between the petrophysical properties density, seismic velocity, heat generation and mineralogical constitution. Earth and Planetary Science Letters 57, 367–376.

Rybach, L., 1986. Amount and significance of radioactive heat sources in sediments. In: Burrus, J. (Ed.), Thermal Modelling in Sedimentary Basins. Editions Techniches, pp. 311–322.

Rybach, L., 1988. Determination of heat production rate. In: Haenel, R., Rybach, L., Stegena, L. (Eds.), Handbook of Terrestrial Heat-Flow Density Determination. Kluwer Academic Press, pp. 125–142.

Sandiford, M., McLaren, S., 2006. Thermo-mechanical controls on heat production distributions and the long-term evolution of the continents. In: Brown, M., Rushmer, T. (Eds.), Evolution and Differentiation of the Continental Crust. Cambridge University Press, pp. 67–91.

Sawyer, E.W., 1996. Melt segregation and magmaflow in migmatites: implications for the generation of granite magmas. Transactions of the Royal Society of Edinburg: Earth Sciences 87, 85–94.

Schott, B., Schmeling, H., 1998. Delamination and detachment of a lithospheric root. Tectonophysics 296, 225–247.

Sun, S., 1980. Lead isotopic studies of young volcanic rocks from mid-ocean ridges, ocean islands and island arcs. Philosophical Transactions of the Royal Society 297, 409–445.

Sun, S., McDonough, W.F., 1989. Chemical and isotopic systematics of ocean basalts: implications for mantle composition and processes. In: Saunders, A.D., Norry, M.J. (Eds.), Magmatism in the Ocean Basins: Geological Society Special Publication, vol. 42, pp. 313–345.

Taylor, S.R., McLennan, S.M., 1985. The Continental Crust: Its Composition and Evolution. Blackwell Scientific Publications, p. 312.

Taylor, S.R., McLennnan, S.M., 1995. The geochemical evolution of the continental crust. Reviews of Geophysics 33, 241–265.

Turcotte, D.L., Schubert, G., 2002. Geodynamics 2nd Edition. Cambridge University Press, Cambridge. (456 pp).
Van Schmus, W.R., 1995. Natural radioactivity of the crust and mantle. Global Earth Physics. A Handbook of Physical Constants (AGU Reference Shelf 1).
Vigneresse, J.L., Barbey, P., Cuney, M., 1996. Rheological transitions during partial melting and crystallization with application to felsic magma segregation and transfer. Journal of Petrology 37, 1579–1600.
Villaseca, C., Downes, H., Pin, C., Barbero, L., 1999. Nature and composition of the lower continental crust in central Spain and the granulite–granite linkage: inferences from granulitic xenoliths. Journal of Petrology 40, 1465–1496.
Webb, P.C., Tindle, A.G., Barrit, S.D., 1987. Factors controlling distribution of heat production in selected UK Granites. Geophysical Research Letters 14, 299–302.
Wedepohl, K.H., 1995. The composition of the continental crust. Geochimica et Cosmochimica Acta 59, 1217–1232.
Whalen, J.B., Currie, K., Chappel, B.W., 1987. A-type granites: geochemical characteristics, discrimination and petrogenesis. Contributions to Mineralogy and Petrology 95, 407–419.
Wilson, M., 1989. Igneous Petrogenesis: Global Appoach. Chapman & Hall, London, p. 466.
Wollenberg, H.A., Smith, A.R., 1987. Radiogenic heat production of crustal rocks: an assesment based on geochemical data. Geophysical Research Letters 14, 295–298.