

Determination of the Crustal and Thermal Structure of the Erzurum-Horasan-Pasinler Basins (Eastern Türkiye) Using Gravity and Magnetic Data

NAFIZ MADEN,¹ ALI AYDIN,² and FAKHRADDIN KADIROV³

Abstract—The Erzurum-Horasan-Pasinler basin, which surrounds Miocene rocks, is oriented approximately E–W and is located in Eastern Anatolia (Türkiye). East Anatolia, where ophiolitic and young volcanic rocks are widespread, is situated in the Alpine–Himalayan fold-thrust fault belt. The NW–SE trending North Anatolian Transform Fault Zone and the NE–SW trending East Anatolian Transform Fault Zone, formed by the compressional regime of East Anatolia, control the main tectonics of the study region. While the Moho and Conrad depths of the study region are 43.0 and 20.9 km, respectively, the average sedimentary thickness has been determined to be 5.2 km by using the power spectrum method. On the other hand, it is found that the depth of the Moho in the region varies from 41.0 to 44.5 km and the depth of the Conrad discontinuity varies between 22 and 26 km, as computed using empirical equations. The basement of the sedimentary layer is calculated to be 6 km by using inversion results applied to the residual gravity data. The Curie point depth and average heat flow value in this region are calculated as 18.0 km and 89.1 m Wm^{−2}, respectively. Geotherm calculations reveal that the Moho temperature is 1,028.0 °C based on the crustal model. The high heat flow values obtained are attributed to tectonic activities and melting of the lithospheric mantle caused by upwelling of the asthenosphere.

Key words: Gravity, Magnetics, Moho, Conrad, Curie, Heat flow, Thermal structure, Erzurum-Pasinler basin, Eastern Türkiye.

1. Introduction

The Eastern Anatolian region is a geologically very complex and intensely deformed high plateau, is seismically active and is a continent–continent collision region located at the collision point of the Arabian and Eurasian Plates within the Alpine–

Himalayan fold-thrust fault belt (ŞENGÖR and KIDD 1979; ŞENGÖR *et al.* 1985; YILMAZ *et al.* 1987; PEARCE *et al.* 1990). For that reason, crust in eastern Türkiye is weak and hot. Focal depths of the earthquakes in this region do not exceed depths of 25 km (REILINGER *et al.* 1997).

In this region, a number of studies (SEBER *et al.* 2001; AL-LAZKI *et al.* 2003; GOK *et al.* 2003; ŞENGÖR *et al.* 2003; ZOR *et al.* 2003; SANDVOL and ZOR 2004; GELİŞLİ and MADEN 2006; ANGUS *et al.* 2006; BARAZ-ANGI *et al.* 2006; PAMUKCU *et al.* 2007; MADEN *et al.* 2009a, b; GÖK *et al.* 2011; MADEN 2012b) have been performed to examine the crustal structure of the Eastern Anatolia and surrounding regions. SEBER *et al.* (2001) found that the crust thickness of the Turkish–Iranian plateau varies between 45 and 50 km. AL-LAZKI *et al.* (2003) and GOK *et al.* (2003) studied P_n tomographic imaging of mantle velocity and anisotropy, and regional S_n wave propagation.

ŞENGÖR *et al.* (2003) determined the crustal thickness of the East Anatolian High Plateau to be only 45 km by using seismic data collected via a network of 29 seismograph stations. ZOR *et al.* (2003) examined the crustal structure of the Anatolian plateau in Eastern Türkiye from teleseismic recordings of a 29-station broadband PASSCAL temporary network using receiver functions. They found that crustal thickness varies from 42 km near the southern part of the Bitlis suture zone to 50 km along North Anatolian fault, and increases from 40 km in the northern Arabian plate to 46–48 km in the middle of the Anatolian plateau.

Tomography studies by SANDVOL and ZOR (2004) showed that the low velocity zone is also underlain by a slightly lower velocity in the upper mantle beneath the northern Arabian plate and the eastern-

¹ Department of Geophysics, Gümüşhane University, Gümüşhane, Türkiye. E-mail: nmaden@gumushane.edu.tr

² Department of Geophysics, Pamukkale University, Denizli, Türkiye.

³ Azerbaijan National Academy of Sciences, Institute of Geology, Baku, Azerbaijan.

most portion of the Anatolian plate. High Bouguer gravity anomalies support the presence of asthenospheric material underlying the Moho in this region (ATEŞ *et al.* 1999). BARAZANGI *et al.* (2006) showed that the Anatolian plateau lithosphere in eastern Türkiye has no lithospheric mantle and computed the average thickness of the crust as 45 km. ANGUS *et al.* (2006) observed three teleseismic S-to-P converted crustal phases which are the Moho with depths ranging between 30 and 55 km, pointing to variable tectonic regimes within this continental collision zone. PAMUKCU *et al.* (2007) evaluated that the Moho depths increase from 38 to 52 km in the Eastern Anatolia region.

MADEN *et al.* (2009a) determined the average Moho, Conrad and basement depth of the eastern Pontides as 35.7, 26.5 and 4.6 km, respectively. Also, they showed that the Moho depths fluctuate between 33.9 and 42.6 km, as obtained from the power spectrum method, and increase from 30.1 to 43.8 km, as determined from gravity inversion studies. MADEN *et al.* (2009b) investigated the tectonic and crustal structure of the eastern Pontides orogenic belt and concluded that the Moho depth varies from 29.0 km in the north to 47.2 km in the south by using the power spectrum method applied to the gravity data. GÖK *et al.* (2011) indicated that the Moho depth reaches 52 km in the southeastern part of the eastern Pontides orogenic belt and northern Armenia. MADEN (2012b) estimated the Moho thickens from north (27.2 km) to south (40.8 km) along the central Pontides.

Many researchers (ATEŞ *et al.* 2005; AYDIN *et al.* 2005; DOLMAZ *et al.* 2005; BEKTAŞ *et al.* 2007; MADEN *et al.* 2009b; MADEN 2009, 2010, 2013) published maps of Curie point depths of the Türkiye mainland and Black sea. AYDIN *et al.* (2005) presented a Curie point isotherm map of all of Türkiye from aeromagnetic data with a spectral analysis technique. While the maximum Curie point depths are between 20–29 km in the Eastern Pontides and Western Taurus belt, the minimum Curie point depths in the Central Anatolian and Aegean region are 6–10 km. DOLMAZ *et al.* (2005) determined the Curie point depth for Western Anatolia ranges between 8.2 and 19.9 km. Curie point depths of central Anatolia vary from 7.9 and 22.6 km. The shallowest Curie point

depths (7.9 km) were observed around the Cappadocia and Erciyes volcanic complexes in Central Anatolia (ATEŞ *et al.* 2005). BEKTAŞ *et al.* (2007) showed Curie point depths in the Eastern Anatolia region vary from 12.9 to 22.6 km. MADEN (2009) estimated the depth to Curie point surface thickens from south (14.8 km) to north (21.8 km) along the central Pontides and determined the average Curie point depth as 20.4 km. MADEN *et al.* (2009b) found that the Curie point depths fluctuate between 14.3 km in the south and 27.9 km in the north by using magnetic data of the Eastern Pontides. MADEN (2010) computed two Curie point depths as 11.4 and 13.7 km, reflecting the high geothermal potential of the Cappadocia and Erciyes volcanic region. MADEN (2013) determined the mean Curie point depth value of 25 km, which is thinner than the Moho depth value in the continental areas.

The first heat flow map for Türkiye was prepared by TEZCAN and TURGAY (1989) with constant thermal conductivity data gathered from exploration wells. TEZCAN (1995) revised this map by using temperatures measured in oil and coal wells mostly drilled in the southeastern Türkiye and Thrace basin. Additionally, İLKİŞİK (1992) made a heat flow map of Türkiye based on the silica and the chemical content of thermal springs. In this map, the average heat flow values are estimated in the western, the central, and the eastern Anatolia as 110.72, 102.78, and 112.81 m Wm^{-2} , respectively. Later, TEZCAN (1995) stated that the heat flow values over the Anatolia should be much higher than those over the Mediterranean and the Black Sea.

MADEN (2009, 2010, 2012a, b, 2013) studied the heat flow values for the Black sea, Eastern Pontides, Central Pontides, and Central Anatolia regions. In the Central Pontides, the heat flow values computed from the Curie point depth values vary from 94.1 m Wm^{-2} in the south (back arc) to 63.9 m Wm^{-2} in the north (arc), indicating that the southern part of the investigated region has a relatively high geothermal potential with respect to the northern part of the region (MADEN 2009). MADEN (2010) reported heat flow values in the Erciyes region of 88.8 and 106.5 m Wm^{-2} based on silica data and conventional heat flow methods, respectively. MADEN (2012a) determined that the surface heat flow values vary between

66.5 and 104.7 m Wm⁻² and computed Moho temperature as 590 ± 60 °C, indicating the presence of a brittle-ductile transition zone along the southern Black sea coast. MADEN (2012b) revealed that the Moho temperatures decrease from 992 °C in the south (back-arc) to 415 °C in the north (arc) of the Central Pontides. MADEN (2012b) also suggested that the Eurasia plate had moved from north to south under the Anatolia plate along the south Black sea coast. More recently, MADEN (2013) determined the Moho temperature increases from 367 °C in the trench (Black sea) to 978 °C in the arc region (Eastern Pontides). The heat flow values at the Moho surface change between 16.4 m Wm⁻² in the Black sea basin and 56.9 m Wm⁻² in the Eastern Pontides.

In this study, we used two different methods to calculate the crustal thickness of the region, the power spectrum method and the empirical equation method. In addition, we evaluated the Curie point depth, thermal gradient, and surface heat flow values of the Erzurum-Horasan-Pasinler to propound the thermal structure of the study region.

2. General Geology and Tectonics

Eastern Anatolia, where the thermal activity is high, is a continent–continent collision region situated in the Alpine–Himalayan fold-thrust fault belt. It is observed in many parts of East Anatolia that ophiolitic and young volcanic rocks are widespread. Two systems of strike-slip faults control the main tectonics of the Plateau. These are NW–SE trending dextral strike-slip faults paralleling the North Anatolian Transform Fault Zone (NATFZ) and NE–SW trending sinistral strike-slip faults paralleling the East Anatolian Transform Fault Zone (EATFZ). In response to the compressional regime of East Anatolia, the North and East Anatolian fault systems were formed in the region (ZOR *et al.* 2003; KOÇYIĞIT *et al.* 2001).

The Erzurum-Horasan-Pasinler basin, which surrounds Miocene rocks, is oriented E–W and located in Eastern Anatolia (Türkiye). The Pontides bound this basin in the north and the Bitlis Mountains are to the south (Fig. 1). The sedimentary sequence in the region is Eocene to Quaternary. The ophiolite and ophiolitic mélangé from the Cretaceous age in the

Palandöken and Kop mountains form the basement of the sedimentary sequence. These sedimentary patches overlapping unconformably with the Late Cretaceous–Middle Eocene are covered with Upper Oligocene–Lower and Middle Miocene sedimentary sequences. In the epoch between the Upper Miocene and the present time, the youngest sedimentary patches of the studied area have been developed. Within this, patches of sedimentary sequences, accumulated since the Upper Oligocene era, play important roles as source beds, reservoir beds, and covered beds. In addition, due to oil seepage and geological data indicating hydrocarbons, geophysical studies have gradually increased in the region (PELIN *et al.* 1980).

According to ŞENGÖR and YILMAZ (1981), in the late Paleocene along the Pontide arc, the Neotethys was subducted beneath the southern margin of Eurasia and was completely consumed. However, recent geologic and geophysical studies (ŞENGÖR *et al.* 2003) showed that, during the middle Miocene, a small branch of the Neotethys subducted northward beneath eastern Türkiye. Extensive melting and collisional volcanism formed the regional uplift in the east Anatolian plateau (YILMAZ 1993; ANGUS *et al.* 2006).

3. Data and Methods

The gravity data used, provided by the General Directorate of Mineral Research and Exploration (MTA) of Türkiye, was acquired at 500 m intervals and with a precision of 0.01 mGal. Gravity values were tied to MTA and General Command of Mapping base stations related to the Potsdam 9,81,260.00 mGal absolute gravity value, which is accepted by the International Union of Geodesy and Geophysics in 1971. Latitude correction was applied according to the 1967 International Gravity Formula. Bouguer reduction and topographic correction to a distance of 167 km were calculated, assuming a terrain density of 2.67 gcm⁻³. A gravity anomaly map of the region is shown in Fig. 1.

The aeromagnetic data gathered between the years 1978 and 1989 were surveyed by the MTA with a 600-m flight altitude above the ground surface. Flight lines vary between one and five kilometers depending on the expected mineral, geothermal, or

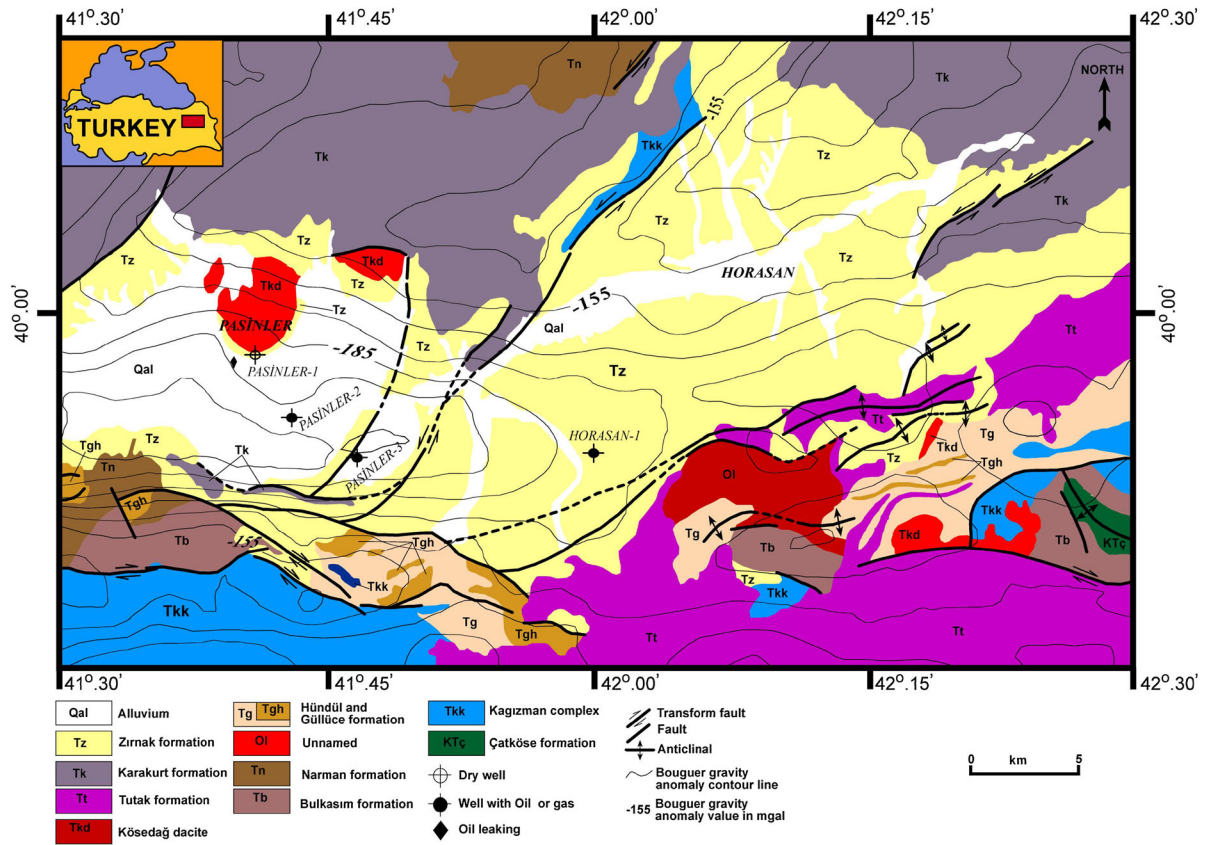


Figure 1

Simplified geological and Bouguer anomaly map of the study region (GELİŞLİ and MADEN 2006). The contour interval of the Bouguer anomaly map is 10 mGal

other resource in the region; sample spacing along the flight lines was approximately 70 m. The flight direction was perpendicular to the regional geologic formations and tectonic structures. The IGRF (1982.5) was removed from the data for preparing the aeromagnetic anomaly map of Türkiye. The year 1982 was chosen for the removal of IGRF since it was the mid-date of the survey span and the larger sectors were overflown during this year. The total component of the geomagnetic field was measured along N–S trending profiles, then measured values were reduced to 1982 with daily variation and direction error corrections. The aeromagnetic anomaly map (Fig. 2.) grid interval of 2.5 km used in this study was obtained from the MTA (AYDIN and KARAT 1995; ATEŞ *et al.* 1999; AYDIN *et al.* 2005).

Spectral analysis, a statistical-approaching method, has been widely used in order to determine

the depth of anomaly sources (BHATTACHARYA 1965, 1966; SPECTOR and BHATTACHARYA 1966; NAIDU 1970; SPECTOR and GRANT 1970; TREITEL *et al.* 1971; MISHRA and NAIDU 1974; CURTIS and JAIN 1975; CIANCIARA and MARCAK 1976; HAHN *et al.* 1976; BLAKELY 1995; HOFSTETTER *et al.* 2000; MADEN and GELİŞLİ 2001; RIVERO *et al.* 2002; BÜYÜKSARAÇ *et al.* 2005; BANSAL *et al.* 2006; GELİŞLİ and MADEN 2006; BEKTAŞ *et al.* 2007; MADEN 2009; MADEN *et al.* 2009a, b, 2010, 2012a). This method provides a relationship between anomaly spectrums and the depth of sources in the wavenumber domain. This method does not require a priori knowledge about the geometry and density contrasts of the source bodies. SPECTOR and GRANT (1970) have shown a simple relation between the power spectrum and depth of prismatic magnetic sources. DIMITRIADIS *et al.* (1987) generalized the application of the spectral statistical method to gravity data observed at the

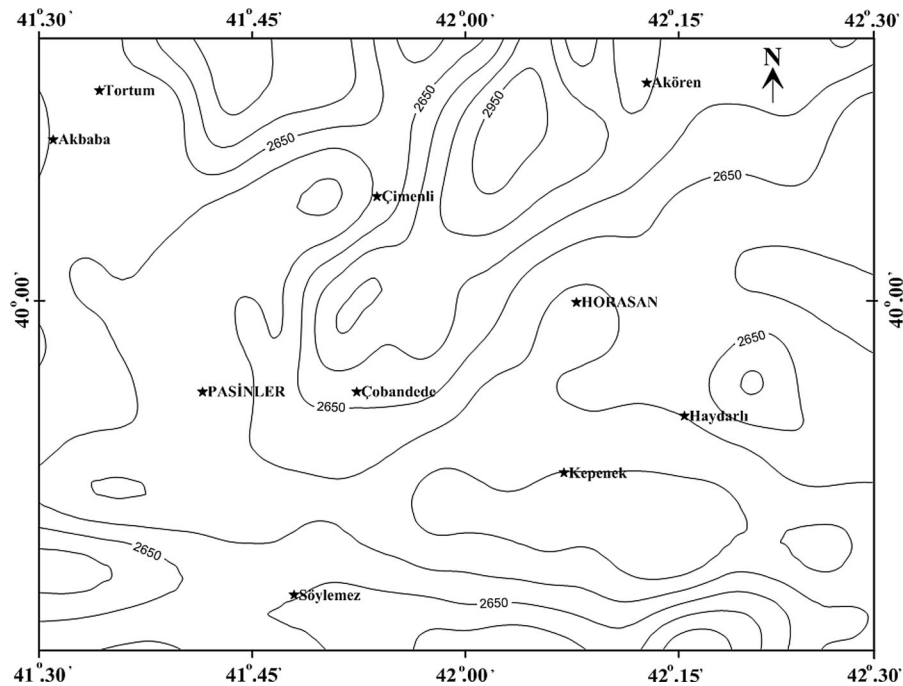


Figure 2

Low-pass filtered map of magnetic anomaly data of the study region for computing Curie-point depths. The contour interval is 50 nT

surface for estimating the mean depth of ensembles of magnetic sources (TREITEL *et al.* 1971). Gravity data is transformed into the wavenumber domain to analyze the wavenumber content of the data. A plot of the logarithm power spectrum versus wavenumber usually shows several straight-line segments that decrease in slope with increasing wavenumber (Fig. 3). The slopes of various segments provide an estimate of the depths to the various levels of anomalous sources. The slope of the low wavenumber parts of the power spectrum curve yield the deep sources, while the slope of the high wavenumber region imply the depth to the shallow gravity sources. The cut-off wavenumber separates the power spectrum curve of the Bouguer gravity data into high and low wavelength. The high and low wavenumber domains are associated with shallow and deep structures, called residual and regional anomalies, respectively.

Following PAL *et al.* (1979), the depth to a given ensemble of prismatic bodies d , can be estimated from the power spectrum of the associated anomaly, by applying:

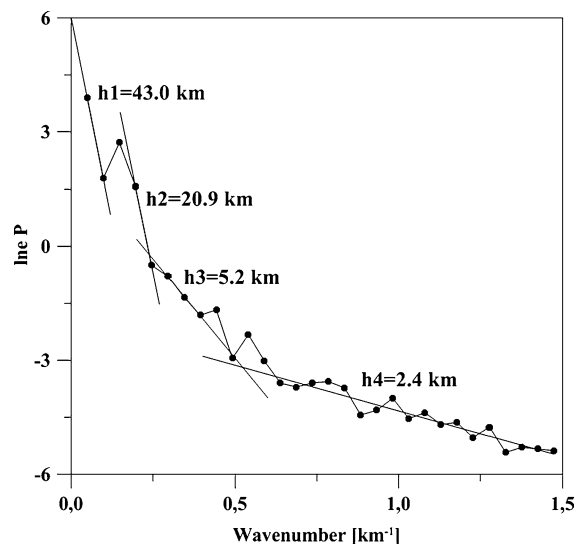


Figure 3

Power spectrum of the Bouguer gravity data of the study region (GELİŞLİ and MADEN 2006). It is a plot of the logarithm of the energy versus wavenumber of the gravity data collected in the Erzurum-Horasan-Pasinler basins and surroundings. Estimate h_1 represents the Moho depth; h_2 , h_3 , and h_4 represent the depth estimates for the main physical interfaces within the crust

$$E(k, 0) = E(0, 0)e^{\left(\frac{2\pi}{N}d - m\right)k}, \quad (1)$$

where k is the wavenumber, m is the slope of the average spectrum, and N is the number of gravity data. The depth (d) to the top of the basement can be easily calculated, since the gradient of $\left(\frac{2\pi}{N}d - m\right)$ should be zero. This technique is a fast and affordable method for estimating the depth of disturbing anomaly sources and gives reliable depth estimates (NWOGBO 1998). When the power spectrum method is applied to the gravity data, the cut off wavenumber is fixed in regional-residual separation and, when applied to the magnetic data, the Curie depth can be estimated (MADEN and GELİŞLİ 2001).

Several empirical studies on crustal structure, as determined from the relation between a Bouguer gravity anomaly (Δg) and seismically-determined crustal thickness (H) performed by several authors (WOLLARD 1959; WOLLARD and STRANGE 1962; RAM BABU 1997; HOFSTETTER *et al.* 2000; RIVERO *et al.* 2002), have shown that they are linearly related. A brief account of the theory on the linear relationship between Δg and H is described below.

Let H_1 and H_2 be the Moho depths at two locations P_1 and P_2 , respectively, with corresponding gravity anomalies Δg_1 and Δg_2 (Fig. 4). Let σ_c and σ_m be the average densities of the crust and upper mantle, respectively. Since $H_2 - H_1$, H_1 and $H_2 \ll x$, where x is the horizontal extent of the

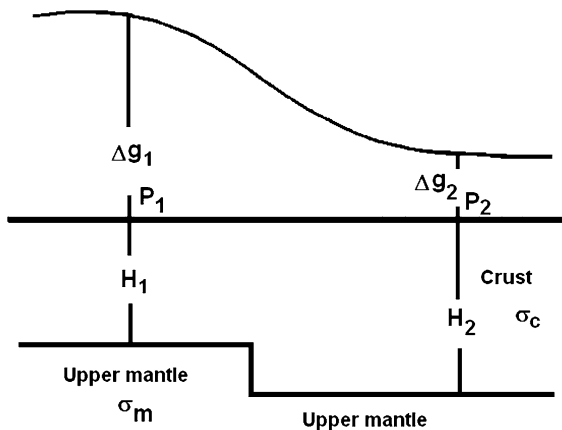


Figure 4

Schematic cross section of a gravity anomaly profile over a density surface (after RAM BABU 1997)

structure involved, the maximum gravity anomaly caused by the plate of thickness $H_2 - H_1$ is given by.

$$\Delta g_1 - \Delta g_2 = 2\pi G(\sigma_m - \sigma_c)(H_2 - H_1), \quad (2)$$

where G is the universal gravitation constant. It follows from Eq. (2) that

$$H_1 = H_2 + M(\Delta g_2 - \Delta g_1), \quad (3)$$

where $M = 1/2\pi G(\sigma_m - \sigma_c)$. If H_0 is the Moho depth corresponding to $\Delta g = 0$, Eq. (3) can be generalized as

$$H = H_0 + M\Delta g, \quad (4)$$

A number of deep seismic sounding profiles in the study region must be available to calculate the Moho depths from Eq. (4). This equation represents a straight line with slope M . The M and H_0 constants are determined from graphs of seismic Moho depths against Bouguer values observed in the same stations (RAM BABU 1997). RAM BABU (1997) showed that the empirical relationships between a Bouguer anomaly and seismically-determined crustal thickness are linearly related. In this study, we used

$$H = 32 - 0.08\Delta g \quad (5)$$

and

$$H_c = 18.6 - 0.031\Delta g, \quad (6)$$

formulas (WOLLARD 1959; DEMENITSKAYA 1967) to estimate the Moho and Conrad depths of the study area. In these formulas, Δg , H , and H_c are gravity anomaly, the Moho depth, and the Conrad depth, respectively. Because the empirical relations are computed in distinct regions, it is important to consider the geology and tectonics of the region if one decides to use the relationship for the gravity data.

In general, sedimentary basins are associated with low gravity values due to low-density sediment fill (ATHY 1930; HOWELL *et al.* 1966). Densities of sedimentary rocks increase with depth mainly due to compaction. For the interpretation of the sedimentary basin gravity anomaly, the generally stacked prism model of BOTT (1960) is used as an initial model. In BOTT'S (1960) method of gravity interpretation, the mathematical geometries of a sedimentary basin were modelled with a series of juxtaposed vertical prisms having equal widths. In the absence of the parameters

of such basins, MURTHY and RAO (1989), LEÃO *et al.* (1996), and BARBOSA *et al.* (1997, 1999) developed methods using uniform density contrast to trace the basement interfaces from the observed gravity anomalies. On the other hand, in order to determine the sedimentary basin configuration, CORDELL (1973) used the exponential depth-density relation, RAO (1990) proposed a quadratic density function, CHAI and HINZE (1988) developed an iterative algorithm in the frequency domain using an exponential density function, and VISWESWARA RAO *et al.* (1993, 1994) utilized the parabolic density function. CHAKRAVARTHI *et al.* (2002) and CHAKRAVARTHI and SUNDARAJAN (2004) represented the sedimentary basins with a series of 3D-juxtaposed vertical prisms whose upper surface is the earth surface and bottom surface is at the bottom of the basin, and whose density contrast is constant.

We apply the gravity inversion method to the residual gravity data of the study region to find out the basement depth of the sedimentary layer in the study area. The initial depth of the sedimentary layer is calculated by using BOTT'S (1960) method. The BOTT (1960) method of gravity interpretation involves an estimation of a sedimentary basin by a series of 2D-juxtaposed blocks. The inversion results of the last iteration that perform the best fit between the observed and calculated gravity values have been used for the sedimentary basement model of the region.

SPECTOR and GRANT (1970) examined statistical properties of magnetic anomaly patterns to obtain the average depths to the top of magnetized bodies. This method gives a connection between the 2D FFT power spectrum of the magnetic anomalies and the depth of magnetic sources by transforming the spatial data into the wavenumber domain. BHATTACHARYYA and LEU (1975, 1977) and OKUBO *et al.* (1985, 1989) provided a two-step method to define the bottom depth (Z_b) of the deepest magnetic sources using the spectral analysis method designed by SPECTOR and GRANT (1970). The first step is the computation of the depth to the centroid of a magnetic body (Z_0) from the slope of the longest wavelength part of the spectrum. The second step is estimation of the top boundary (Z_t) of the magnetized source from the slope of the second longest wavelength spectral segment. The basal depth, inferred from the Curie

point depth, Z_b is then obtained from these two depth values by using the formula: $Z_b = 2Z_0 - Z_t$ (TANAKA *et al.* 1999; OKUBO *et al.* 1985).

Heat flow can be computed using the formula

$$q = k \left[\frac{\theta}{Z_b} \right], \quad (7)$$

where q is the heat flow in m Wm^{-2} , k is the coefficient of thermal conductivity in $\text{Wm}^{-1} \text{K}^{-1}$, θ is the Curie temperature in $^{\circ}\text{C}$, and Z_b is the Curie point depth in km. In this equation, it is assumed that the surface temperature is 0°C and no heat sources exist between the Earth's surface and the Curie point depth. In this equation, the Curie point depth is inversely proportional to the heat flow (TURCOTTE and SCHUBERT 1982; TANAKA *et al.* 1999; STAMPOLIDIS *et al.* 2005).

On the other hand, surface heat flow density for the region can be computed from the equation of SHARMA *et al.* (2005) as follows:

$$Q_s = 640H^{-0.68}, \quad (8)$$

where Q_s is the heat flow density in m Wm^{-2} and H is the Curie temperature depth in km. Typical surface heat flow densities are about $30\text{--}40 \text{ m Wm}^{-2}$ in fore-arc regions, more than 100 m Wm^{-2} in the volcanic front, and about $70\text{--}80 \text{ m Wm}^{-2}$ in back-arc regions (HYNDMAN and LEWIS 1999). The thermal structure of the crust can be approximated by steady state solution of the heat conduction equation as follows:

$$\nabla^2 T = -\frac{A}{k}, \quad (9)$$

where T is the temperature ($^{\circ}\text{C}$), A is the radioactive heat production in $\mu \text{ Wm}^{-3}$, and k is the thermal conductivity in $\text{Wm}^{-1} \text{K}^{-1}$. In case of constant heat production, the analytical solution to the heat equation with depth can be given with the following equation of CERMAK *et al.* (1991):

$$T(z) = T_0 + \frac{Q_s}{k} z - \frac{A_0}{2k} z^2, \quad (10)$$

where T is the temperature at depth z in $^{\circ}\text{C}$, T_0 is the annual mean temperature at the Earth's surface (5.6°C obtained from Turkish State Meteorological Service for the region), k is the thermal conductivity in $\text{Wm}^{-1} \text{K}^{-1}$, and A_0 is the heat production value in

$\mu \text{ Wm}^{-3}$ at the surface. HE *et al.* (2009) used the heat production values for the upper, middle, and lower crusts and the mantle as 1.10, 0.83, 0.37, and 0.24 $\mu \text{ Wm}^{-3}$, respectively. In addition, JOKINEN and KUKKONEN (1999) used the heat production values for the upper, middle, and lower crust and the lithospheric mantle as 1.8, 0.6, 0.2, and 0.002 $\mu \text{ Wm}^{-3}$, respectively.

TURCOTTE and SCHUBERT (1982) suggested that the geothermal conductivity (k) of basalts and granites are 1.3–2.9 and 2.4–3.8 $\text{Wm}^{-1} \text{K}^{-1}$, respectively. TEZCAN (1979) assigned a geothermal conductivity value of 2.1 $\text{Wm}^{-1} \text{K}^{-1}$ to Neogene clayey formations. RAO *et al.* (1970) suggested using a thermal conductivity value for the sediments of 3.14 $\text{Wm}^{-1} \text{K}^{-1}$. CORREIA and JONES (1995) used the thermal conductivity values of the upper, middle, and lower crust as 2.7, 2.5, and 2.1 $\text{Wm}^{-1} \text{K}^{-1}$, respectively. RIMI (1999) suggested that the thermal conductivity value of the lower crust is 2.6 $\text{Wm}^{-1} \text{K}^{-1}$. In this study, the thermal conductivity values of the upper and middle crust are taken from CORREIA and JONES (1995) and RIMI (1999) as 2.7 and 2.6 $\text{Wm}^{-1} \text{K}^{-1}$, respectively.

4. Crustal Structure of the Erzurum-Horasan-Pasinler Basins

The Bouguer gravity map of the Erzurum-Horasan-Pasinler basin shows closed anomalies because of the low-density sedimentary units. Especially, the Pasinler basin produces large negative gravity anomalies (Fig. 1). It is shown that Bouguer gravity anomaly contours lay parallel to the main tectonic elements of the region. The high frequency magnetic anomalies near the Pasinler, Çobandede, and Çimenli region appear to be associated with volcanic rocks (Fig. 2). Regional geological features, topography of the region, and magnetic core fields produce long wavelength magnetic anomalies and could affect the centroid depth estimates and probably create erratic estimates of the Curie point depths (OKUBO *et al.* 1985). For that reason, the magnetic data of the research area were reduced to the pole and were low-pass filtered to prevent the effects of topography.

The power spectrum method (SPECTOR and GRANT 1970) has been performed by taking a 2D Fourier transformation of the Bouguer gravity data of the region. To compute the average depth component of the gravity sources in the study area, logarithmic amplitude spectra of the radial wavenumber were drawn (Fig. 3). In this graphic, power spectrum analysis reveals three distinct linear segment lines that give three source depths. The first one, which is the low wavenumber region between 0 and 0.1 km^{-1} and respects contributing long wavelength sources, is considered due to density difference in the Moho discontinuity (43.0 km). The second one, which is an intermediate region in the interval 0.15–0.25 km^{-1} , is associated with the density difference between the upper and lower crust called the Conrad interface (20.9 km). Sediment filling (5.2 km) cause the third region to have a high wavenumber between 0.3–0.5 km^{-1} . The last one, which is the larger wavenumbers region (0.5–1.0 km^{-1}), is considered to be the youngest sedimentary patche (2.4 km).

Various empirical relations are generally based on knowledge of the Moho depth through seismic methods. These empirical relationships between the Bouguer anomaly (Δg) and seismically-determined crustal thickness (H) reveal that they are linearly related (WOLLARD 1959; WOLLARD and STRANGE 1962; RAM BABU 1997; HOFSTETTER *et al.* 2000; RIVERO *et al.* 2002). In this study, we employ the Eqs. (5) and (6) suggested by WOLLARD (1959) and DEMENITSKAYA (1967) to estimate the Moho and Conrad depths of the study region, respectively. Eq. (5) proposed by WOLLARD (1959) was suggested for the whole earth instead of a specific region. However, the depths computed by the Wollard equation give, in general, the information on the crustal thickness in Eastern Türkiye. In addition, it provides an opportunity to compare the results obtained from other methods. By using the Wollard equation, we fixed and mapped the Moho depth of the study region. In this map, the Moho depth ranges between 41.0 and 44.5 km (Fig. 5). In the same region, we estimate the average Moho depth as 43.0. When we take into account two depth values, there is almost a 2.0 km difference between the power spectrum method and the Wollard equation. On the other hand, the depth values of the Conrad discontinuities are determined to vary

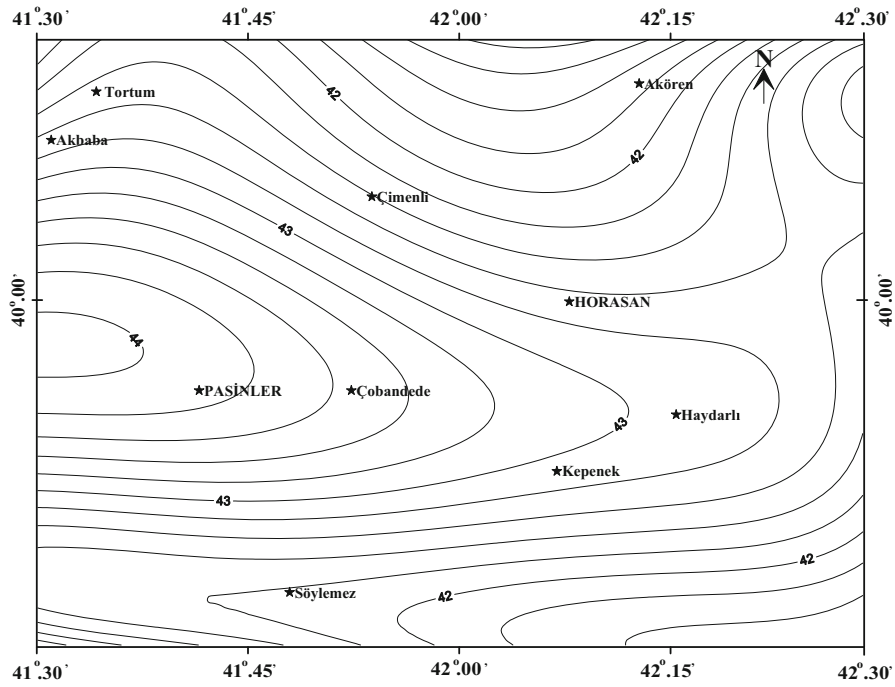


Figure 5

Moho depth of the study area calculated from the regional gravity anomaly by using empirical equations. Contour interval is 0.5 km

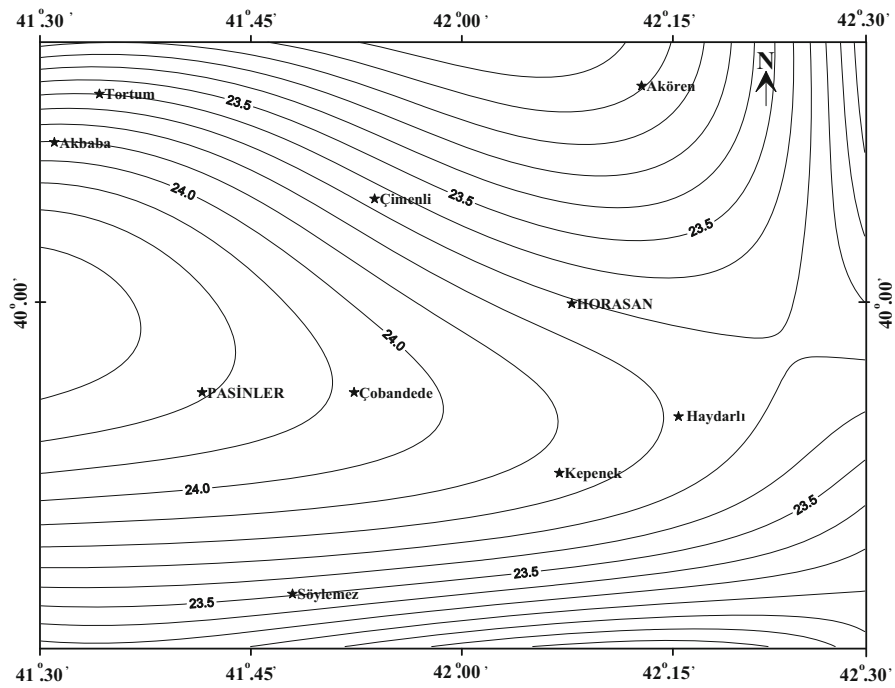


Figure 6

Conrad depth of the study area calculated from the regional gravity anomaly by using empirical equations. Contour interval is 0.2 km

between 22 and 26 km (Fig. 6) by using Eq. (6) (DEMENTITSKAYA 1967). In the same way, the maximum difference is approximately 5.1 km between the power spectrum method and empirical equations.

We used a 3D gravity inversion method to estimate the basement depth of the Erzurum-Horasan-Pasinler sedimentary basins. Initial depths of the sedimentary basin are derived from the Bouguer slab formula and subsequently updated iteratively based on the differences between the observed and theoretical gravity anomalies until the modeled gravity anomalies resemble the observed ones. The residual gravity data shown in Fig. 7a was computed by subtracting the regional trend surface from the Bouguer gravity field. The calculated gravity field and the basement relief maps of the basin are shown in Fig. 7b and c, respectively. The closeness of fit between the observed (Fig. 7a) and calculated (Fig. 7b) gravity anomalies shows the validity of the results. The maximum basement depth of the study region is found to be 6 km by using 3D gravity inversion of the residual gravity data (Fig. 7c). This depth value coincides fairly well with the basement depth value (5.2 km) determined from power spectrum curves. The basement depths of sedimentary layer counters are parallel to the tectonic units and maps obtained for Moho (Fig. 5) and Conrad (Fig. 6) discontinuities.

5. Thermal Regime of the Region

The Curie point depth estimates the average depth of magnetic sources and it is believed they reflect the thermal structure of the region. Partial melting of the lower crust is thought to have contributed to the present thermal structure of the crust. Previous studies show that the Curie point depth varies greatly according to the geological context. It is shown that the Curie point depths are shallower than about 10 km within volcanic and geothermal areas, 15–25 km in island arcs and ridges, deeper than 20 km in plateau regions, and deeper than 30 km in trench areas (TANAKA *et al.* 1999).

In this paper, we utilize the available aeromagnetic data over the Erzurum-Horasan-Pasinler region to calculate the Curie isotherm depths. Curie point

depth estimation requires the deepest magnetic sources. However, reduced-to-pole magnetic anomaly data contain both long and small wavelength anomalies. It is important to take into account the fact that the magnetic data must be reduced-to-the-pole before low pass filtering to remove the effects of topography (TANAKA *et al.* 1999). The low pass filtered magnetic anomaly map (Fig. 2) was used for the Curie point depth estimation.

We determined the depth to the tops and centroids of the magnetic sources from an azimuthally-averaged power spectrum of the low pass filtered magnetic anomaly data to be 9.6 and 13.8 km, respectively (Fig. 8a, b). The average Curie point depth is then computed from those two depth values as 18.0 km for this region. Bektaş *et al.* (2007) calculated the Curie point depths vary between 17 and 19 km in Eastern Anatolia, near Erzurum. The geothermal gradient ($\frac{dT}{dz}$) between the Earth's surface and the Curie point depth (Z_b) can be calculated by using the formula $\frac{580}{Z_b}$ giving $32.2 \text{ } ^\circ\text{C km}^{-1}$ in the region. In this calculation, the Curie point temperature was assumed to be $580 \text{ } ^\circ\text{C}$.

Typical surface heat flow density values are about $30\text{--}40 \text{ m Wm}^{-2}$ in fore-arc regions, more than 100 m Wm^{-2} in the volcanic front, and about $70\text{--}80 \text{ m Wm}^{-2}$ in back-arc regions (HYNDMAN and LEWIS, 1999). Two surface heat flow density values are estimated as 90.3 and 87.9 m Wm^{-2} from the Curie temperature depth of the region by using the Eqs. (7) and (8), as suggested by SHARMA *et al.* (2005). The average heat flow value for these values is 89.1 m Wm^{-2} .

POLLACK and CHAPMAN (1977) empirically established that 60 % of the surface heat flow density is attributed the mantle heat flow density, whereas 40 % is the contribution of radiogenic sources in the upper crust. Hence, 53.5 m Wm^{-2} is the contribution from the mantle and the remainder (35.6 m Wm^{-2}) is from the crust. High mantle heat flow density values (53.5 m Wm^{-2}) should be related to the melting, and collisional volcanism formed the regional uplift in the east Anatolian plateau. as suggested by YILMAZ (1993) and ANGUS *et al.* (2006).

Geotherm calculations reveal that the Moho temperature (Fig. 9) is $1,028.0 \text{ } ^\circ\text{C}$ based on the crustal model, as determined from power spectral

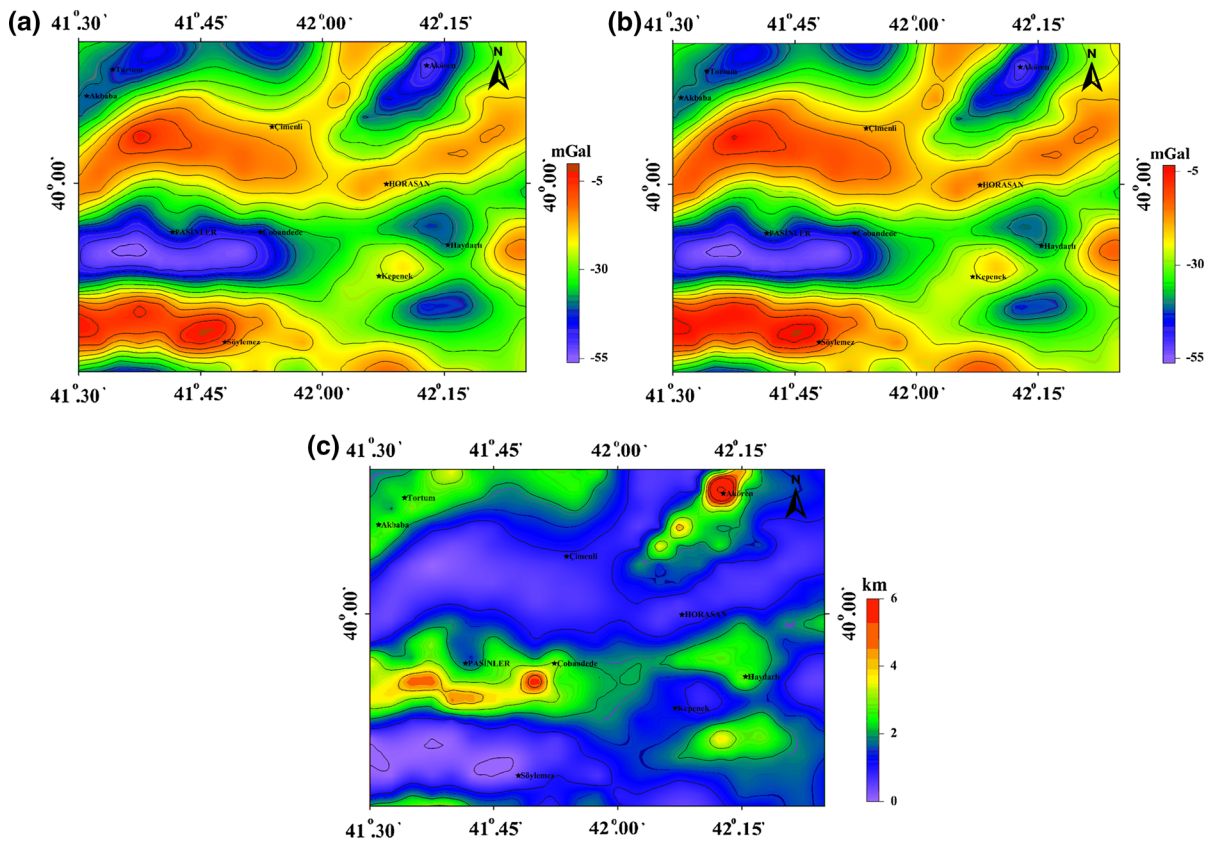


Figure 7

Observed (a) and calculated (b) gravity fields, and basement relief (c) maps of the Erzurum-Horasan-Pasinler basins. Basement relief map of the region determined from observed residual gravity using a constant density contrast of -0.45 gcm^{-3} . Basement relief map contour interval is 0.5 km. Contour interval of the observed and calculated residual gravity maps is 5 mGal

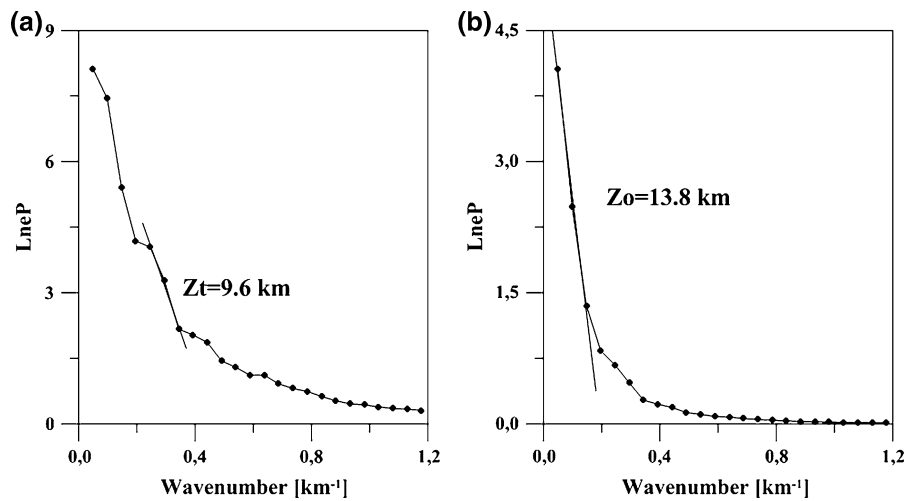


Figure 8

Power spectrum curve of the aeromagnetic data. a top boundary of magnetic layer (Z_t); b centroid of a magnetic layer (Z_0) derived from the slope of the power spectrum curve (GELİŞLİ and MADEN 2006)

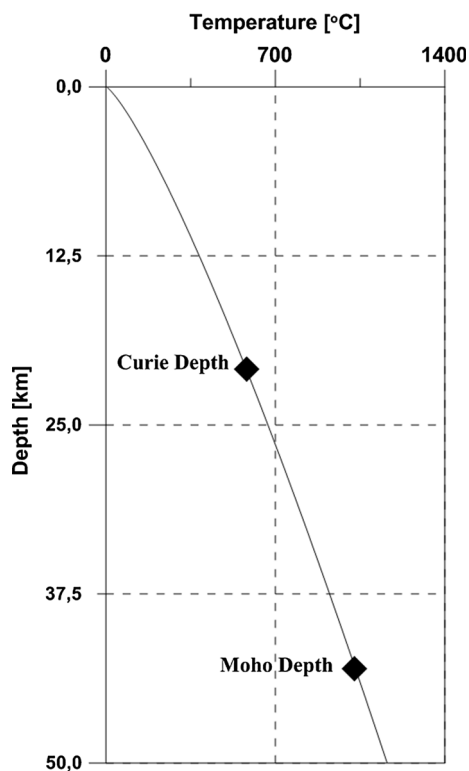


Figure 9

Temperature as a function of depth for the Erzurum-Horasan-Pasinler basins based on spectrum analysis studies

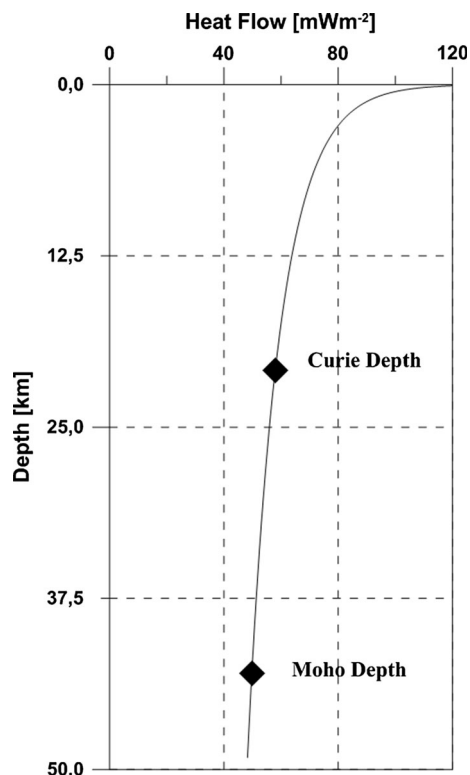


Figure 10

Heat flow versus depth profile for the Erzurum-Horasan-Pasinler basins. Crustal thickness values are obtained from power spectral analysis

analysis for the study region. High heat flow values might be related to melting of the lithospheric mantle caused by upwelling of the asthenosphere. The high heat flow density values are attributed to the tectonic activities, as suggested by ŞENGÖR and YILMAZ (1981) and BOZKURT (2001). The depth of Curie temperature is found to be 21 km, which is more or less consistent with the Curie temperature depth value given by BEKTAŞ *et al.* (2007) and GELİŞLİ and MADEN (2006). In addition, this depth coincides fairly well with the Curie point depth estimated using the power spectrum method (18 km). The geothermal gradient value can be computed by using this depth value as 27.7 °C km^{-1} . The difference between the geothermal gradient values obtained using the power spectrum method and the geotherm calculation is estimated as 4.5 °C km^{-1} . The mantle heat flow value (Fig. 10) determined from the geotherm calculations is 48.7 m Wm^{-2} in the Erzurum-Horasan-Pasinler basin. Table 1 shows the parameters used in the geotherm calculation for the region.

Table 1

Depth values for the crustal model, thermal conductivity, and heat production values used for the geotherm calculation

Layer	Depth (km)	Thermal conductivity ($\text{Wm}^{-1}\text{ K}^{-1}$)	Heat production ($\mu\text{ Wm}^{-3}$)
Sediments	5	2.1 ^a	2.93 ^c
Upper crust	21	2.7 ^b	1.10 ^f
Lower crust	43	2.6 ^c	0.37 ^f
Mantle		4.2 ^d	0.002 ^d

^a TEZCAN (1979)

^b CORREIA and JONES (1995)

^c RIMI (1999)

^d JOKINEN and KUKKONEN (1999)

^e MADEN (2012a)

^f HE *et al.* (2009)

6. Discussions

We applied the power spectrum method to Bouguer gravity data and found the Moho, Conrad, and

basement of the sedimentary layer as 43.0, 20.9, and 5.2 km, respectively. On the other hand, by applying the empirical equation of WOLLARD (1959), we determined the Moho depth values vary between 41.0 and 44.5 km. It is shown that these values derived from the power spectrum method and empirical equations are consistent with the results computed by SEBER *et al.* (2001), AL-LAZKI *et al.* (2003), GOK *et al.* (2003), ZOR *et al.* (2003), SANDVOL and ZOR (2004), and ANGUS *et al.* (2006). PAMUKCU *et al.* (2007) estimated a Moho depth of 47 km near the Erzurum-Horasan-Pasinler region. This depth is only 2.5 km different from our Moho depth value computed using the Wollard equation.

TANAKA *et al.* (1999) showed that Curie point depths are shallower than about 10 km in volcanic and geothermal areas, 15–25 km at island arcs and ridges, deeper than 20 km at plateaus, and deeper than 30 km at trenches. We determined the Curie point depth to be 18.0 km in this region by using the power spectrum method applied to the magnetic data. On the other hand, the Curie point depth value for the region is computed as 21 km by using geotherm computations. This study show that the maximum difference for the Curie point depth values between this study and AYDIN *et al.* (2005) is 2 km in the study region. In the Eastern Anatolia region, BEKTAŞ *et al.* (2007) found that the Curie point depth values change from 12.9 to 22.6 km, conforming reasonably well with the results obtained in this study. In addition, our study results are not too dissimilar from the results of TANAKA *et al.* (1999) found in the plateaus.

We determined the maximum depth of the sedimentary basin basement in the research area to be 6 km by using the gravity inversion method; using the power spectrum method, the basement depth was calculated to be 5.2 km. The difference between the power spectrum method and inversion results is 0.8 km.

In the Erzurum-Horasan-Pasinler basin, the Moho temperature is found to be 1,028.0 °C, based on the crustal model determined via power spectral analysis applied to the gravity data. The mantle heat flow value in the study region is estimated as 48.7 m Wm⁻². The average Curie point depth for the region, as determined from a spectral analysis technique applied to magnetic data, was found to be 18.0 km. The depth of the Curie temperature (21 km) obtained from

geotherm calculations corresponds to the crustal magma chambers, as previously identified by KESKIN *et al.* (1998, 2006), BEKTAŞ *et al.* (2007), KARSLI *et al.* (2007), EYUBOGLU *et al.* (2011a, b), ÖZDEMİR *et al.* (2011), and MOGHADAM *et al.* (2014). The thermal gradient values are 32.2 and 27.7 °C km⁻¹. Surface heat flow value is estimated at 89.1 m Wm⁻².

The high heat flow value (89.1 m Wm⁻²) found in the east Anatolian plateau is accompanied by the values estimated by HYNDMAN and LEWIS (1999). Thin magnetic crust found in this region accounts for tectonically active regions also associated with higher heat flow values. It is also possible that partial melting of the crust caused from the magma rising contributed to the current high geothermal potential in the study area.

7. Conclusions

The following conclusions have been drawn from the present study:

1. Moho depth values change between 41.0 and 44.5 km by using empirical equations for the region.
2. The Moho, Conrad, and sedimentary basement depths are estimated as 43.0, 20.9, and 5.2 km, respectively, as determined using the power spectrum method.
3. The maximum depth of the sedimentary basin basement is 6 km, as derived by using the 3D gravity inversion method.
4. The Curie point depth and the geothermal gradient values are computed as 18.0 km and 32.2 °C km⁻¹, respectively, by using the power spectrum method.
5. The Curie point depth and the geothermal gradient values are 21.0 km and 27.7 °C km⁻¹, respectively, as determined via geotherm computations.
6. The average surface heat flow density value is evaluated as 89.1 m Wm⁻².
7. The Moho temperature and mantle heat flow values in the Erzurum-Horasan-Pasinler basin are 1,028.0 °C and 48.7 m Wm⁻², respectively.
8. High heat flow values are possibly related to crustal magma chambers, the ophiolitic and young volcanic rocks of which are observed in many

parts of Eastern Anatolia continent–continent collision region situated in the Alpine–Himalayan fold-thrust fault belt; the high heat values might also be related to melting of the lithospheric mantle caused by upwelling of the asthenosphere.

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REFERENCES

- AL-LAZKI, A., SANDVOL, E., SEBER, D., TURKELLI, N., MOHAMAD, R. and BARAZANGI, M., 2003. *Tomographic Pn velocity and anisotropy structure beneath the Anatolian plateau (Eastern Turkey) and the surrounding regions*, Geophys. Res. Lett., 30, 8040.
- ANGUS, D. A., WILSON, D. C., SANDVOL, E. and NI, J. F., 2006. *Lithospheric structure of the Arabian and Eurasian in eastern Turkey from S-wave receiver functions*, Geophys. J. Int., 166, 1335–1346.
- ATEŞ, A., KEAREY, P. and TUFAN, S., 1999. *New Gravity and Magnetic Maps of Turkey*, Geophys. J. Int. 136, 499–502.
- ATEŞ, A., BILIM, F. and BÜYÜKSARAC, A., 2005. *Curie point depth investigation of Central Anatolia, Turkey*, Pure and Applied Geophysics 162, 357–371.
- ATHY, L. F., 1930. *Density, porosity and compaction of sedimentary rocks*, Bull. Assoc. Pet. Geol., 14, 1–24.
- AYDIN, I. and KARAT, H. I., 1995. *A general view of Türkiye aeromagnetic maps*, Jeofizik 9, 41–44 (in Turkish).
- AYDIN, I., KARAT, H. I., and KOCAK, A., 2005. *Curie point depth map of Turkey*, Geophys. J. Int. 162, 633–640.
- BANSAL, A. R., DIMRI V. P. and SAGAR, G. V., 2006. *Depth Estimation from Gravity Data Using the Maximum Entropy Method (MEM) and the Multi Taper Method (MTM)*, Pure and Applied Geophysics 163, 1417–1434.
- BARAZANGI, M., SANDVOL, E., and SEBER, D., 2006. *Structure and tectonic evolution of the Anatolian plateau in eastern Turkey*, in Dilek, Y., and Pavlides, S., eds., *Post-collisional tectonics and magmatism in the Mediterranean region and Asia*: Geological Society of America Special Paper 409, 463–474.
- BARBOSA, V. C. F., SILVA, J. B. C. and MEDEIROS, W. E., 1997. *Gravity inversion of basement relief using approximate equality constraints on depths*, Geophysics 62, 1745–1757.
- BARBOSA, V. C. F., SILVA, J. B. C. and MEDEIROS, W. E., 1999. *Gravity inversion of a discontinuous relief stabilized by weighted smoothness constraints on depth*, Geophysics 64, 1429–1437.
- BEKTAŞ, O., RAVAT, D., BUYUKSARAC, A., BILIM, F., and ATEŞ, A., 2007. *Regional Geothermal Characterisation of East Anatolia from Aeromagnetic, Heat Flow and Gravity Data*, Pure and Applied Geophysics 164, 975–998.
- BLAKELY, R. J., 1995. *Potential theory in gravity and magnetic applications*: Cambridge University Press.
- BHATTACHARYA, B. K., 1965. *Two Dimensional Harmonic Analysis as a tool magnetic Interpretation*, Geophysics 30, 829–857.
- BHATTACHARYA, B. K., 1966. *Continuous spectrum of the total magnetic anomaly due to a rectangular prismatic body*, Geophysics 31, 97–121.
- BHATTACHARYYA, B. K. and LEU, L. K., 1975. *Spectral analysis of gravity and magnetic anomalies due to two-dimensional structures*, Geophysics 40, 993–1013.
- BHATTACHARYYA, B. K. and LEU, L. K., 1977. *Spectral Analysis of gravity and magnetic anomalies due to rectangular prismatic bodies*, Geophysics 42, 41–50.
- BOTT, M. P. H., 1960, *The use of rapid digital computing methods for direct gravity interpretation of sedimentary basins*, Geophys. J. R. Astron. Soc. 3 pp. 63–67.
- BOZKURT, E., 2001. *Neotectonics of Turkey—a synthesis*, Geodinamica Acta 14, 3–30.
- BÜYÜKSARAC, A., JORDANOVA, D., ATEŞ, A., and KARLOUKOVSKI, V., 2005. *Interpretation of the Gravity and Magnetic Anomalies of the Cappadocia Region, Central Turkey*, Pure and Applied Geophysics 162, 2197–2213.
- CERMAK, V., BODRI, L. and RYBACH, L. 1991. *Radioactive heat production in the continental crust and its depth dependence*. In: Cermak, V. and Rybach, L. (eds.), *Terrestrial Heat Flow and the Lithosphere Structure* (Springer-Verlag, New York, 1991) pp. 23–69.
- CHAI, Y. and HINZE, W. J., 1988. *Gravity inversion of an interface above which the density contrast varies exponentially with depth*, Geophysics 53, 837–845.
- CHAKRAVARTHI, V., RAGHURAM, H. M. and SINGH, S. B., 2002. *3-D forward gravity modelling of basement interfaces above which the density contrast varies continuously with depth*, Computers and Geosciences, 28, 53–57.
- CHAKRAVARTHI, V. and SUNDARARAJAN, N., 2004. *Automatic 3-D gravity modelling of sedimentary basins with density contrast varying parabolically with depth*, Computers and Geosciences, 30, 601–607.
- CIANCIARA, B. and MARCAK, H., 1976. *Interpretation of Gravity Anomalies by Means of Local Power Spectra*, Geophysical Prospecting 24, 273–286.
- CORDELL, L., 1973. *Gravity analysis using an exponential density-depth function-San Jacinto Graben: California*, Geophysics, 38, 684–690.
- CORREIA, A. and JONES, F. W., 1995. *A magnetotelluric survey in a reported geothermal area in southern Portugal*, Proc. World Geothermal Congr. 2, 927–931.
- CURTIS, C. E. and JAIN, S., 1975. *Determination of Volcanic Thickness and Underlying Structures from Aeromagnetic Maps of the Silet Area of Algeria*, Geophysic 40(1), 79–90.
- DEMENITSKAYA, R. M., 1967. *Crust and Mantle of the Earth*, Nedra, Moscow, 288 p.
- DIMITRIADIS, K., TSELENTIS, G. A. and THANASSOULAS, K., 1987. *A BASIC program for 2-D spectral analysis of gravity data and source-depth estimation*, Computers and Geosciences 13, 549–560.
- DOLMAZ, M. N., HISARLI, Z. M., USTAÖMER, T. and ORBAY, N., 2005. *Curie point depths based on spectrum analysis of aeromagnetic data, West Anatolian Extensional Province, Turkey*, Pure and Applied Geophysics 162, 571–590.
- EYUBOĞLU, Y., SANTOSH, M. and CHUNG, S. L., 2011a, *Crystal fractionation of adakitic magmas in the crust-mantle transition zone: Petrology, geochemistry and U-Pb zircon chronology of the Seme adakites, Eastern Pontides, NE Turkey*, Lithos 121, 151–166.

- EYUBOĞLU, Y., SANTOSH, M. and CHUNG, S. L., 2011b, *Petrochemistry and U-Pb ages of adakitic intrusions from the Pulur massif (Eastern Pontides, NE Turkey): Implications for slab roll-back and ridge subduction associated with Cenozoic convergent tectonics in eastern Mediterranean*. *Journal of Geology* 119, 394–417.
- GELİŞLİ, K. and MADEN, N., 2006, *Analysis of Potential Field Analysis in Pasinler-Horasan Basin, Eastern Turkey*, *Journal of the Balkan Geophysical Society*, 9, 1, 1–7.
- GÖK, R., SANDVOL, E., TURKELLI, N., SEBER, D. and BARAZANGI, M., 2003, *Sn attenuation in the Anatolian and Iranian plateaus and surrounding regions*, *Geophys. Res. Lett.*, 30, 8042.
- GÖK, R., MELLORS, R. J., SANDVOL, E., PASYANOS, M., HAUK, T., TAKEDATSU, R., YETİRMİŞLİ, G., TEOMAN, U., TURKELLI, N., GODOLADZE, T. and JAVAKISHVILİ, Z., 2011, *Lithospheric velocity structure of the Anatolian plateau-Caucasus-Caspian region*, *J. Geophys. Res.* 116, B05303.
- HAHN, A., KING, E. G. and MISHRA, D. C., 1976, *Depth Estimation of Magnetic Sources by Means of Fourier Amplitude Spectra*, *Geophysical Prospecting* 24, 287–308.
- HE, L., HU, S., YANG, W. and WANG, J., 2009, *Radiogenic heat production in the lithosphere of Sulu ultrahigh-pressure metamorphic belt*. *Earth and Planetary Science Letters* 277, 525–538.
- HOFSTETTER, A., DORBATH, C., RYBAKOV, M. and GOLDSCHMIDT, V., 2000, *Crustal and upper mantle structure across the Dead Sea rift and Israel from teleseismic P-wave tomography and gravity data*, *Tectonophysics* 327, 37–59.
- HOWELL, L. G., HEINTZ, K. O. and BARRY, A., 1966, *The development and use of a high precision downhole gravity meter*, *Geophysics* 31, 764–772.
- HYNDMAN, R. D., LEWIS, T. J. 1999, *Geophysical consequences of the Cordillera-Craton thermal transition in southwestern Canada*. *Tectonophysics* 306, 397–422.
- İLKİŞİK, O. M., 1992, *Silica heat flow estimates and lithospheric temperature in Anatolia*, *Proceedings, XI. Congress of World Hydrothermal Organization*, İstanbul, Mayıs 13–18, pp. 92–104.
- JOKINEN, J. and KUKKONEN, I. T., 1999, *Random modelling of the lithospheric thermal regime: forward simulations applied to uncertainty analysis*. *Tectonophysics* 306, 277–292.
- KARSLI, O., CHEN, B., AYDIN, F. and ŞEN, C., 2007, *Geochemical and Sr-Nd-Pb isotopic compositions of the Eocene Dölek and Sariçiçek plutons, Eastern Turkey: implications for magma interaction in the genesis of high-K calc-alkaline granitoids in a post-collision extensional setting*, *Lithos* 98, 67–96.
- KESKIN M., PEARCE J. A. and MITCHELL J. G., 1998, *Volcano-stratigraphy and geochemistry of collision-related volcanism on the Erzurum-Kars plateau, northeastern turkey*, *Journal of Volcanology and Geothermal Research* 85, 355–404.
- KESKIN M., PEARCE J. A., KEMPTON P. D. and GREENWOOD P., 2006, *Magma-crust interactions and magma plumbing in a post collisional setting: geochemical evidence from the Erzurum-Kars volcanic plateau, eastern turkey*, In: *Postcollisional tectonics and magmatism in the Mediterranean region and Asia*. Eds., Dilek Y., Pavlides S. The Geological Society of America Special Paper 409, 475–505.
- KOCYIGIT, A., YILMAZ, A., ADAMIA, S. and KULOSHVILI, S., 2001, *Neotectonics of East Anatolian Plateau (Turkey) and Lesser Caucasus: implications for transition from thrusting to strike-slip faulting*, *Geodinamica Acta*, 14, 177–195.
- LEÃO, J. W. D., MENEZES, P. T. L., BELTRÃO, J. F. and SILVA, J. B. C., 1996, *Gravity inversion of basement relief constrained by the knowledge of depth at isolated points*. *Geophysics* 61, 1702–1714.
- MADEN, N. and GELİŞLİ, K., 2001, *Evaluation of the Bouguer Gravity Anomalies of the Erzurum-Horasan-Pasinler area with the power spectrum method*, 13th International Petroleum Congress and Exhibition of Turkey, Proceedings, p. 139–147, Ankara-Turkey.
- MADEN, N., 2009, *Crustal thermal properties deduced from spectral analysis of magnetic data in Central Pontides (Turkey)*. *Turkish Journal of Earth Sciences* 18, 383–392.
- MADEN, N., GELİŞLİ, K., BEKTAŞ, O. and EYÜBOĞLU, Y., 2009a, *Two-and-three-dimensional crust topography of the Eastern Pontides (NE TURKEY)*. *Turkish Journal of Earth Sciences* 18, 225–238.
- MADEN, N., GELİŞLİ, K., EYÜBOĞLU, Y. and BEKTAŞ, O., 2009b, *Determination of tectonic and crustal structure of the Eastern Pontide Orogenic Belt (NE Turkey)*. *Pure and Applied Geophysics* 166, 1987–2006.
- MADEN, N., 2010, *Curie-point depth from spectral analysis of magnetic data in Erciyes stratovolcano (Central TURKEY)*, *Pure and Applied Geophysics* 167, 349–358.
- MADEN, N., 2012a, *One Dimensional thermal modeling of the Eastern Pontides Orogenic Belt (NE TURKEY)*, *Pure and Applied Geophysics* 169, 235–248.
- MADEN, N., 2012b, *Two-dimensional geothermal modeling along the Central Pontides magmatic arc (Northern Turkey)*, *Survey in Geophysics*, 33, 275–292.
- MADEN, N., 2013, *Geothermal Structure of the Eastern Black Sea Basin and the Eastern Pontides Orogenic Belt: Implications for Subduction Polarity of Tethys Oceanic Lithosphere*. *Geoscience Frontiers* 4(4), 389–398.
- MISHRA, D. C. and NAIDU, P. S., 1974, *Two-Dimensional Power Spectral Analysis of Aeromagnetic Fields Using Fast Digital Fourier Transform Techniques*, *Geophysical Prospecting* 27, 344–361.
- MOGHADAM, H. S., GHORBANI, G., KHEDR, M. Z., FAZLİA, N., CHIRADIA, M., EYUBOĞLU, Y., SANTOSH, M., FRANCISCO, C. G., LOPEZ, M., GOURGAUD, A. and ARAI, S., 2014, *Late Miocene K-rich volcanism in the Eslamieh Peninsula (Saray), NW Iran: Implications for geodynamic evolution of the Turkish-Iranian High Plateau*. *Gondwana Research*, 26, 1028–1050.
- MURTHY, I. V. R. and RAO, S. J., 1989, *A Fortran 77 program for inverting gravity anomalies of two-dimensional basement structures*, *Computers and geosciences* 15, 1149–1156.
- NAIDU, P. S., 1970, *Statistical Structure of Aeromagnetic Field*, *Geophysics* 35, 279–292.
- NWOGBO, P. O., 1998, *Spectral prediction of magnetic sources depths from simple numerical models*, *Computers and Geosciences*, 24, 9, 847–852.
- OKUBO, Y., GRAF, J. R., HANSEN, R. O., OGAWA, K., and TSU, H., 1985, *Curie Point Depths of the Island of Kyushu and Surrounding Areas, Japan*, *Geophysics* 53, 481–494.
- OKUBO Y., TSU H. and OGAWA K., 1989, *Estimation of Curie point temperature and geothermal structure of island arcs of Japan*. *Tectonophysics* 159, 279–290.
- Özdemir, Y., Blundy, J. and Güleç, N., 2011, *The importance of fractional crystallization and magma mixing in controlling chemical differentiation at Süphan stratovolcano, eastern Anatolia, Turkey*, *Contributions to Mineralogy and Petrology* 162, 573–597.
- PAL, P. C., KHURANA, K. K. and UNNIKISHNA, P., 1979, *Two examples of spectral approach to source depth determination in gravity and magnetics*, *Pure and Applied Geophysics* 117, 772–783.

- PAMUKCU, O. A., AKÇİĞ, Z., DEMİRBAŞ, Ş. and ZOR, E., 2007. *Investigation of Crustal Thickness in Eastern Anatolia Using Gravity, Magnetic and Topographic Data*, Pure and Applied Geophysics 11, 1420–9136.
- PEARCE, J. A., BENDER, J. F., DE LONG, S. E., KIDD, W. S. F., LOW, P. J., GÜNER, Y., ŞAROĞLU, F., YILMAZ, Y., MOORBATH, S. and MITCHELL, J. G., 1990. Genesis of collision volcanism in Eastern Anatolia, Turkey. In: Le Fort, J. A. Pearce and A. Pecher (Editors), *Collision Magmatism*, J. Volcanol. Geotherm. Res. 44, 184–229.
- PELIN, S., ÖZSAYAR T., GEDİK, I. and TOKEL, S., 1980. *Geological investigation of Pasinler (Erzurum) Basin for oil*, MTA 729, 25–40.
- POLLACK, H. N. and CHAPMAN, D. S., 1977. *On the regional variation of heat flow, geotherms and lithosphere thickness*. Tectonophysics 38, 279–296.
- RAM BABU, H. V., 1997. *Average crustal density of the Indian lithosphere: an inference from gravity anomalies and deep seismic soundings*, Journal of Geodynamics, 23, 1–4.
- RAO, D. B. 1990. *Analysis of gravity anomalies of sedimentary basins by an asymmetrical trapezoidal model with quadratic density function*, Geophysics 55, 226–231.
- RAO, R. U. M., VERMA, R. K. and GUPTA, M. L., 1970. *Heat flow at Dam and Mohapani, Satpura Gondwana basin India*, Earth Planet. Sci. Lett. 7, 406–412.
- REILINGER, R. E., MCCLUSKY, S. C., ORAL, M. B., KING, R. W., TOKSOZ, M. N., BARKA, A. A., KINIK, I., LENK, O. and SANLI, I., 1997. *Global Positioning System measurements of present-day crustal movements in the Arabia-Africa-Eurasia plate collision zone*. Journal of Geophysical Research 102(B5), 9983–9999.
- RIMI, A., 1999. *Mantle heat flow and geotherms for the main geologic domains in Morocco*. Int. Journ. Earth Sciences 88, 458–466.
- RIVERO, L., PINTO, V. and CASAS, A., 2002. *Moho depth structure of the eastern part of the Pyrenean belt derived from gravity data*. J. Geodyn. 33, 315–332.
- SANDVOL, E. and ZOR, E., 2004. Upper mantle P and S-wave velocity structure beneath eastern Anatolian plateau. EOS, Trans. Am. Geophys. Un., AGU Fall Meeting 2004, S13B–1056.
- SEBER, D., SANDVOL, E., BRINDISI, C. and BARAZANGI, M., 2001. *Crustal model for the Middle East and North Africa region: implications for the isostatic compensation mechanism*, Geophys. J. Int., 147, 630–638.
- SHARMA, S. R., RAO, V. K., MALL, D. M. and GOWD, T. N., 2005. *Geothermal structure in a seismoactive region of Central India*, Pure and Applied Geophysics 162, 129–144.
- SPECTOR, A. and BHATTACHARYA, B. K., 1966. *Energy Spectrum and Autocorrelation Functions of Anomalies Because of Simple Magnetic Models*, Geophysical Prospecting 14, 242–272.
- SPECTOR, A. and GRANT, F. S., 1970. *Statistical models for interpreting aeromagnetic data*, Geophysics 35, 293–302.
- STAMPOLIDIS, A., KANE, I., TSOKAS, G. N. and TSOURLOS, P., 2005. *Curie point depths of Albania inferred from ground total field magnetic data*. Surveys in Geophysics 26, 461–480.
- ŞENGÖR, A. M. C. and KIDD, W. S. F., 1979. *Post-collisional tectonics of the Turkish-Iranian plateau and a comparison with Tibet*, Tectonophysics 55, 361–376.
- ŞENGÖR, A. M. C. and YILMAZ, Y., 1981. *Tethyan evolution of Turkey: a plate tectonic approach*, Tectonophysics, 75, 181–241.
- ŞENGÖR, A. M. C., GÖRÜR, N. and ŞAROĞLU, F., 1985. Strike-slip faulting and related basin formation in zones of tectonic escape: Turkey as a case study. In: Biddle, K. T., Christie-Blick, N. (Eds.), *Strike-slip deformation, basin formation and sedimentation*. Society of Economy, Palaeontology and Mineralogy Special Publication 37, 227–264.
- ŞENGÖR, A. M. C., ÖZEREN, S., GENÇ, T. and ZOR, E., 2003. *East Anatolian high plateau as a mantle supported, north-south shortened domal structure*, Geophys. Res. Lett., 30(24).
- TANAKA, A., OKUBO, Y., and MATSUBAYASHI, O., 1999. *Curie-point depth based on spectrum analysis of the magnetic anomaly data in East and Southeast Asia*, Tectonophysics. 306, 461–470.
- TEZCAN, A. K., 1979. Geothermal studies, their present status and contribution to heat flow contouring in Turkey. In *Terrestrial Heat Flow in Europe* (eds. Cermak, V. and Rybach, L.), Springer, Berlin, pp. 283–291.
- TEZCAN, A. K., 1995. Geothermal explorations and heat flow in Turkey, in: *Terrestrial Heat Flow and Geothermal Energy in Asia*. Edited by: Gupta, M. L. and Yamano, M., Science Publishers, Lebanon, New Hampshire, 23–42.
- TEZCAN, A. K. and TURGAY, M. I., 1989. Heat flow map of Turkey, General Directorate of Mineral Research and Exploration (MTA), Department of Geophysics Research, Ankara (in Turkish, unpublished).
- TREITEL, S., CLEMENT, W. G. and KAUL, R. K., 1971. *The Spectral Determination of Depths to Buried Magnetic Basement Rocks*, Geophys. Jour. Roy. Astr. Soc. 24, 415–428.
- TURCOTTE, D. L. and SCHUBERT, G., 1982. *Geodynamics Applications of Continuum Physics to Geologic Problems*. Wiley, New York, 450 pp.
- VISWESWARA RAO, C., CHAKRAVARTHI, V. and RAJU, M. L., 1993. *Parabolic Density function in sedimentary basin modelling*, Pure and Applied Geophysics 3, 493–501.
- VISWESWARA RAO, C., CHAKRAVARTHI, V., and RAJU, M. L., 1994. *Forward modelling: Gravity anomalies of two-dimensional bodies of arbitrary shape with hyperbolic and parabolic density functions*, Computer and Geosciences 20, 873–880.
- WOLLARD, G. P., 1959. *Crustal structure from gravity and seismic soundings*, J. Geophys. Res., 64, 1524–1544.
- WOLLARD, G. P. and STRANGE, W. E., 1962. Gravity anomalies and crust of the earth in the Pacific basin, in: *The crust of the Pacific basin*. Geophysical Monograph 6, 12.
- YILMAZ, Y., ŞAROĞLU, F. and GÜNER, Y., 1987. *Initiation of the neomagnetism in Eastern Anatolia*, Tectonophysics, 134, 177–199.
- YILMAZ Y., 1993. *New evidence and model on the evolution of the southeast Anatolian orogen*. Geol. Soc. Am. Bull. 105, 251–271.
- ZOR, E., SANDVOL, E., GURBUZ, C., TURKELLI, N., SEBER, D. and BARAZANGI, M., 2003. *The crustal structure of the East Anatolian plateau (Turkey) from receiver functions*, Geophys. Res. Lett., 30, 8044.