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Isostatic Compensation in Western Anatolia with Estimate of the Effective Elastic Thickness

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Abstract: Isostasy is concerned with how the crust and mantle adjusts to shifting loads of limited spatial and temporal dimensions. By analysing the frequency content of gravity and topography data, it has been possible to determine the compensation scheme of a region. In this study, the compensation mechanism of the Western Anatolia, which has dynamic loads, was investigated by isostatic response functions. Effective elastic thickness in Western Anatolia region was estimated based on admittance and coherence between gravity and topography data.

The two most reliable indicators of lithospheric strength are the focal depth distribution of earthquakes and relation of gravity anomalies with topography. For this reason, the seismogenic thickness and thermal structure of Western Anatolia was correlated with the effective elastic thickness. The results of this study showed that the strength of the lithosphere of the Western Anatolia resided in average 6 km.

Key Words: effective elastic thickness, isostatic response functions, gravity, Western Anatolia, admittance analysis

Efektif Elastik Kalınlık Kestirimi ile Batı Anadolu'daki İzostatik Dengeleme

Özet: İzostazi, sınırlı uzay ve zaman ortamında yer değiştiren yüklere karşı kabuk ve mantonun nasıl bir tepki verdiği ile ilgilidir. Gravite ve topoğrafya verilerinin frekans içeriğini analiz ederek, bir alana ait dengeleme mekanizması tanımlanabilir. Bu çalışmada, dinamik yüklere sahip Batı Anadolu Bölgesi'nin dengeleme mekanizması, izostatik yanıt fonksiyonları ile incelenmiştir. Batı Anadolu Bölgesi'ndeki efektif elastik kalınlık kestirimi, gravite ve topografya verileri arasındaki girişim (admittance) ve uyuma (coherence) bağlı olarak yapılmıştır.

Deprem odak dağılımı ve gravite ile topografya arasındaki ilişki, litosferik dayanımın en iyi iki göstergesidir. Bu amaçla, sismojenik kalınlık ve Batı Anadolu'nun termal yapısı arasında efektif elastik kalınlık ile ilişki kurulmuştur. Bu çalışmanın sonucunda Batı Anadolu'ya ait litosferik dayanımın ortalama 6 km olduğu saptanmıştır.

Anahtar Sözcükler: Efektif elastik kalınlık, izostatik yanıt fonksiyonları, gravite, Batı Anadolu, girişim analizi

Introduction

The most important point in plate tectonics is the thickness of the lithosphere and how it behaves for long geological eras. With the development of isostasy, one of the basic assumptions of the plate tectonics was put forward which is a fact that the plates have long been remained rigid during large time scales. Briefly, the concept of isostasy is an important milestone in defining the lithosphere.

Nowadays, two models of compensation are commonly accepted. In the first isostatic model, compensation occurs with the thickening under topography (Airy 1855) or with the lateral variances in the density of the crust (Pratt 1855). In the second type of model, known as flexural model, loads are supported by the elastic stresses in the lithospheric plate lying over the fluid and weak astenosphere (Vening Meinesz 1932; Gunn 1943; Walcott 1970). According to this model of flexure, lithosphere is bowed in the regions where large loads act and the crustal thickness is well below the average. As a result of this, stresses are seen in the upper and lower parts of the plate which are exposed to bending. While the stresses form the fragile faultings in the brittle portion of this boundary, on the other hand, the stresses in the lower ductile part lead to flows which easily changes shape (Watts 2001). The boundary where the brittle plate ends and the ductile plate starts, gives us the thickness at which lithosphere is resistant to topographic load, namely the effective elastic thickness.

The response of the plate in the flexural model, in other words the flexural rigidity is characterized by the effective elastic thickness. Similar to the Airy model, topographic loads appear with the bending downward and the crustal thickness. What it differs from the Airy model is that the flexural rigidity is not accepted as zero. In the approach of the interpretation of the isostatic compensation regarding this model, the relation between the gravity anomalies induced by the underground masses and the topography is used. In order to define this relation in the domain wave-number and frequency, and linear transfer function techniques were developed (Dorman & Lewis 1970). By using and developing these techniques, the estimation of effective elastic thickness was achieved in several studies (e.g., McKenzie & Bowin 1976; Zuber *et al*. 1989; Hartley *et al.* 1996; Watts 2001; Rajesh & Mishra 2003; Pamukçu 2004; Yurdakul *et al*. 2005; Pamukçu & Yurdakul 2006; Luis & Neves 2006).

In this study, an approach was developed towards the isostatic model of the lithosphere of the Western Anatolian region where considerably complex tectonic incidents occur. For this purpose, by using the admittance function, the lithospheric flexure model of the region was analysed with the help of Bouguer gravity and topographic data.

To examine the accuracy of the effective elastic thickness estimations, the coherence and the penalty function were computed. Furthermore by using software package (Braitenberg *et al.* 2006; Wienecke 2006) gravity inversion was calculated to estimate the Crust-Mantle interface for using of the effective elastic thickness estimations.

In the final stage of the study, the seismogenic zone in the Western Anatolian region was investigated together with the heat flow, thermal gradient and the effective elastic thickness.

Regional Tectonic Setting of the Western Anatolia

The Western Anatolian region, being within the Alpine-Himalayan orogenic belt, is a part of the extensive compressional zone which lies between Arabian, African and Eurasian plates (Figure 1). It is one of the most tectonically active and rapidly deforming and extending areas in the world (e.g., Dewey & Şengör 1979; Jackson & McKenzie 1984; Şengör *et al.* 1985; Eyidoğan & Jackson 1985; Şengör 1987; Ambraseys 1988; Seyitoğlu & Scott 1991; Taymaz *et al*. 1991; Reilinger *et al.* 1997; Ambraseys & Jackson 1998; Bozkurt 2001).

The fundamental tectonic structures of Western Anatolia are the Sakarya and the İzmir-Ankara Suture Zone, the Menderes Massif, grabens and the Taurides (e.g., Şengör & Yılmaz 1981; Okay *et al.* 1996; Barka & Reilinger 1997; Bozkurt 1996, 2003, 2004, 2007; Emre 1996; Lips *et al.* 2001; Sözbilir 2001, 2002; Yılmaz *et al.* 2000). The Aegean Arc is on the southern side of the area (Taymaz *et al.* 1991; Papazachos *et al.* 2000).

There are two types of basins trending at E–W and NE–SW direction in western Anatolia (see Şengör *et al.* 1985; Şengör 1987; Seyitoğlu & Scott 1991, 1992; Emre 1996; Koçyiğit *et al.* 1999; Bozkurt 2000, 2001, 2003; Yılmaz *et al.* 2000; Sözbilir 2001, 2002, 2005; Bozkurt & Sözbilir 2004; Kaya *et al.* 2004; Tokçaer *et al.* 2005; Aldanmaz 2006; Erkül *et al.* 2006; Emre & Sözbilir 2007 for further reading). The region has been extending in N–S direction since the Early Miocene. Today, in the region approximately 30–40 mm/year extending is occured in N–S direction (Oral *et al.* 1995; Le Pichon *et al.* 1995). The westward movement of Anatolia is seen in the counterclockwise rotation as well as transition (Westaway 1994; McKenzie 1970; Dewey & Şengör 1979; Rotstein 1984; Jackson & Mc Kenzie 1988). According to Bozkurt (2001), this is taken up by a response of the continental lithosphere moving laterally away from zones of compression (tectonic escape), to minimize topographic relief and to avoid subduction of buoyant continental material. There is still a controversy about westward motion in western Turkey (e.g., Dewey & Şengör 1979; McKenzie 1972; Şengör *et al* 1985; Seyitoğlu & Scott 1992; Koçyiğit *et al* 1999; Bozkurt & Sözbilir 2004, 2006).

Extension in Western Anatolia has been attributed to the following tectonic models: (1) *tectonic escape model*: the westward extrusion of the Anatolian block along its boundary structures, North Anatolian Fault and East Anatolian Fault since the Late Serravalian (Dewey & Şengör 1979); (2) *back-arc spreading model*: back-arc extension caused by the south-southwestward migration of the Aegean trench system (McKenzie 1972); (3) *orogenic collapse model*: lateral spreading of the over-thickening crust following the latest Palaeogene collision across Neotethys during the latest Oligocene (Seyitoğlu & Scott 1992); (4) *episodic, two-stage basin formation*: a Miocene–Early Pliocene graben formation (orogenic collapse) followed by a Plio–Quaternary rift-mode stage (westward escape of the Anatolia) under N–S extension

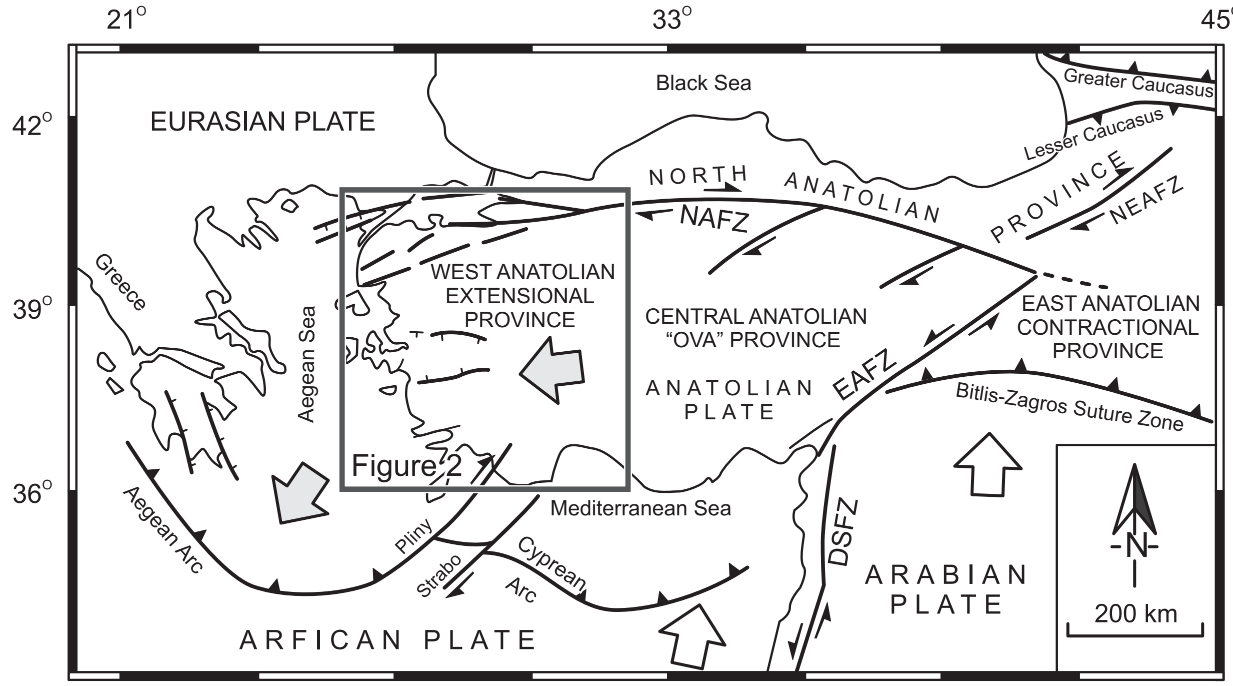

Figure 1. Main features of Turkey (Bozkurt 2001) and location of the study area. K– Karlıova, KM– Kahramanmaraş, DSFZ– Dead Sea Fault Zone, EAFZ– East Anatolian Fault Zone, NAFZ– North Anatolian Fault Zone, NEAFZ– Northeast Anatolian Fault Zone. Heavy lines with half arrows are strike-slip faults with arrows showing relative movement sense. Heavy lines with filled triangles shows major fold and thrust belt: small triangles indicate direction of vergence. Heavy lines with open triangles indicate an active subduction zone, its polarity indicated by the tip of small triangles. The heavy lines with hachures show normal faults: hachures indicate down-thrown side. Open arrows indicate relative movement direction of African and Arabian plates; bold filled arrows, relative motion of Anatolian Plate. Short arrows show the sense of plate motion, half arrows the relative motion senses on strike-slip Faults. The hatched area shows the transition zone between the western Anatolian extensional province and the central Anatolian '*ova*' province.

(e.g., Koçyiğit *et al.* 1999; Bozkurt & Sözbilir 2004, 2006).

Methods

Admittance and Coherence

The approach in the interpretation of the isostatic compensation is achieved by investigating the relation between the gravity anomalies that are induced by underground masses and the topography.

Admittance isostatic response function Z (k), was determined by analyzing the relationship between the Fourier transforms of the gravity and topography (Dorman & Lewis 1970). The relationship is given by

$$G(k) = Z(k)\ .\ T(k) \tag{1}$$

where k (= 2π/wavelength) is wave-number, G(k) and T(k) are discrete Fourier transforms of gravity and topography. Observed admittance is computed by using the cross-spectrum between the gravity and topography and power spectrum of the topography. McKenzie & Bowin (1976) used complex conjugates (*) to eliminate the noise of the data, therefore, defined the observed admittance function as

$$z(k) = \frac{\sum_{r=1}^{N} G_r(k)\,T_r^{\,*}(k)}{\sum_{r=1}^{N} T_r(k)\,T_r^{\,*}(k)} \tag{2}$$

Here, N is the number of data within the profiles that are taken at equal length from the gravity and topography maps that have the same sampling interval.

The profiles in equation (2) were investigated for different tectonic (McKenzie & Bowin 1976) and same or different geological features (Watts 1978). In this study, the four profiles were selected parallel to each other in order to reflect the western Anatolian extensional region in Figure 1 and remaining within the continental part (Figure 2).

Coherence analysis is important since it provides a method for estimating the portion of the spectral domain of gravity which is supposed generated by topography. Within this band, located at intermediate wavelengths, the coherence is approximately 1. Towards short-wavelenghts the coherence is approximately 0 since gravity effect of topography is suppressed by the distance of the source due to depth and smoothing given by gravity filter. In order to investigate the coherence between gravity and topography anomaly McKenzie & Bowin (1976) and Watts (1978) found it useful to compute the coherence:

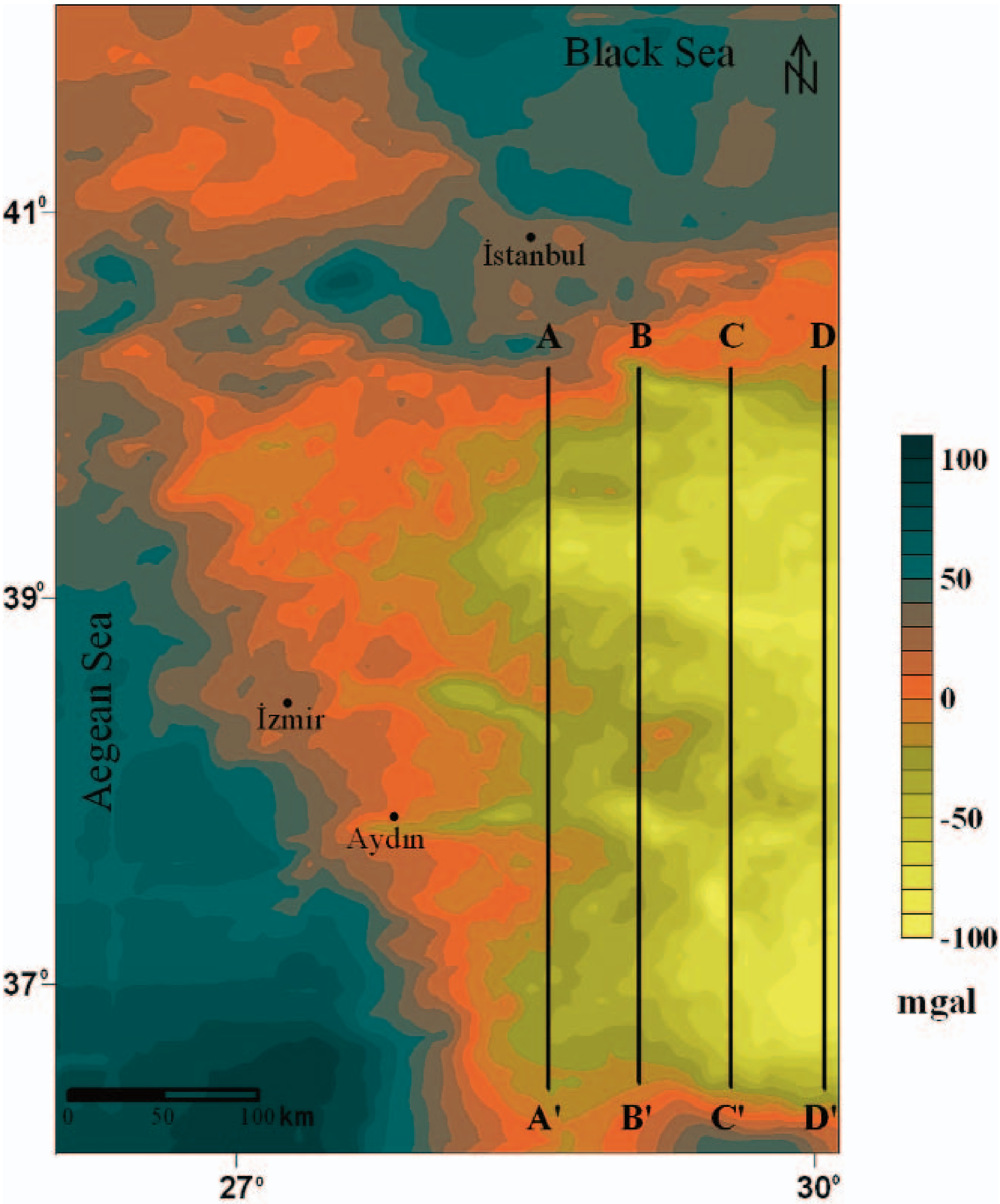

Figure 2. Bouguer gravity map of the field and A–A', B–B', C–C', D–D' profiles.

$$\gamma^2(k) = \frac{C(k)\,C^*(k)}{E_0(k)\,E_1(k)} \tag{3}$$

In the presence of noise, Munk & Cartwright (1966) indicate that a better measure of coherence is given by

$$\gamma_0^2(k) = \frac{(N\gamma^2(k) - 1)}{(N - 1)} \tag{4}$$

Here N is the number of data. Where *C(k)* is cross-spectrum of observed gravity and topography data, $E_0(k)$ and $E_1(k)$ are the power spectra of observed gravity and topography data. They are given by:

$$C(k) = \frac{1}{N}\sum_{r=1}^{N} G_r(k)\,T_r^*(k) \tag{5}$$

$$E_0(k) = \frac{1}{N}\sum_{r=1}^{N} G_r(k)\,G_r^*(k) \tag{6}$$

$$E_1(k) = \frac{1}{N}\sum_{r=1}^{N} T_r(k)\,T_r^*(k) \tag{7}$$

Different isostasic model theories and admittance equations are approached to the relation given in equation (2) which is about the gravity and topography. Generally, theoretical admittance values used for the flexure model were calculated following the method given by McKenzie & Fairhead (1997)

$$Z_c = 2\pi G\rho_c(1 - (\exp(-kt_c))/A$$

$$\text{where } A = 1 + \frac{Dk^4}{g(\rho_m - \rho_c)} \tag{8}$$

$$D = E\,\frac{T_e^3}{12(1 - \sigma^2)}$$

Equation (8) assumes that the density structure of the crust and density of material below the assumed flexed elastic plate. In Equation 8: (1) D is rigidity, (2) k (= 2π/wavelength) is wave-number, (3) G is gravitational constant, (4) E (= 10^{11} Pa) is Young's modulus, (5) σ (= 0,25) is Poisson's ratio, (6) g (= 9,8 m/s^2) is the gravity acceleration, (7) ρ_c (= 2,7 g/cm^3) is average crustal density, (8) ρ_m (= 3,3 g/cm^3) is density of material below the assumed flexed elastic plate, (9) t_c is effective depth of compensation, (10) T_e is the effective elastic thickness, (11) Z_c is theoretical admittance function.

According to McKenzie & Fairhead (1997), Equation (8) expresses that the density structure of the crust and upper mantle can be described by one layer of constant density and variable thickness overlying a constant density half-space. Since the real crustal density is not constant, variations of density within the crust will also produce gravity anomalies. Therefore t_c should be described as an effective depth of compensation. In general, t_c will not be same as the observed crustal thickness. Similar remarks apply to T_e, the effective elastic thickness, where a model with one elastic layer whose properties are constant is also an obvious simplification.

The best fit between the observed and the theoritical admittance values gives the effective elastic thickness (T_e) of the field.

Any method which uses the penalty function (H_f) should provide estimates of T_e value. The minimum value of H_f is defined as goodness of T_e values. The H_f is defined as

$$H_f = \left[\frac{1}{N}\sum\left(\frac{Z_0 - Z_c}{\Delta Z_0}\right)^2\right]^{\frac{1}{2}} \tag{9}$$

where H_f is misfit, Z_0 is the mean value of the observed admittance and ΔZ_0 is the standard deviation of Z_0 in the frequency domain, N is number of data.

Modelling the Crustal-Mantle Interface

In this study software package (Braitenberg *et al.* 2006; Wienecke 2006) which was developed in cooperation between the University of Trieste and the NGU (Geological Survey of Norway), Trondhem was used to calculate the inverse modelling of gravity field. As a result of the calculation, crustal-mantle interface was found.

For calculating the gravity inversion, the reference depth *d* of the density interface and the density contrast across the interface $\Delta\rho$ as starting parametres are required. If $g_0(x,y)$ is the Bouguer gravity field in Cartesian coordinates *x, y* and $g_d(x,y)$ is the downward continued field to the depth *d*, then the Fourier transform of the Bouguer gravity field $FT[g_0]$ can be related to the Fourier transform of the gravity field. $FT[g_d]$ is given by

$$FT[g_d] = e^{d\sqrt{k_x^2 + k_y^2}}\,FT[G_0] \tag{10}$$

where k_x, k_y are the wave numbers along the coordinates axes. It is assumed that the field is generated by a sheet mass located at the depth *d*. The surface density of a this mass $\rho(x,y)$ is given by:

$$\rho(x,y) = \frac{1}{2\pi G} g_d(x,y) = \frac{1}{2\pi G} FT^{-1}[FT[g_d]] \qquad (11)$$

where FT^{1} is the inverse Fourier Transform. The mass that produces the gravity field can be interpreted as a horizontally varying surface density. This can be described with a model of an undulating boundary, which separates two layers with a density contrast $\Delta\rho$ (Braitenberg & Zadro 1999). The undulation amplitude of the boundary is then given by

$$r_1(x,y) = \frac{1}{\Delta\rho}\rho(x,y) \qquad (12)$$

The first approach of gravity field by Parker (1972) was created by the vertical extension limit. A series of rectangular prism was applied to the vertical extension limit by Braitenberg & Zadro (1999) and with the application of the algorithm that was developed by Nagy (1966), $g_1(x,y)$ gravity field was calculated. Residual gravity field is represented with the contrast between observatory field $g_0(x,y)$ and calculated field $g_1(x,y)$. Thus, $\delta g_1(x,y)$, is given by the equation;

$$\delta g_1(x,y) = g_0 - g_1 \qquad (13)$$

Residual field is a way of adjustment and affects on the undulation magnitude of the density boundary. This situation continues iteratively and in every k iteration step, residual gravity field, and undulation magnitude of the boundary is derived $\delta g_k(x,y)$.

Analysis

In the application, Bouguer gravity (from Directorate of Mineral Research and Exploration, MTA, Turkey 1979) and topographic (from Synthetic Aperture Radar, SAR) data of Western Anatolia lying approximately between 27– 30º E longitudes and 36– 40º N latitudes with 5 km sampling intervals were mapped (Figures 2 & 3). Along A–A', B–B', C–C', D–D' profiles shown in Figure 2, sections were taken in N–S direction, with 50 km intervals, 350 km in length.

The methods of estimating T_e that can be used when the coherence between topography and gravity is high. The coherence (Figure 4) which is calculated from equation (4) is high for 0.01<wavenumber<0.1. The admittance which is determined from the spectral analysis (Watts 2001) is also smooth within this waveband (Figure 5). It shows that the waveband must be selected 0.01<wavenumber<0.1 to estimate T_e of the Western Anatolia.

Then, in order to determine the t_c parameter in equation (8), software package was used (Braitenberg *et al.* 2006; Wienecke 2006). By using the software which contains application of inverse solution to the Bouguer gravity data, crust-mantle interface values were calculated (Figure 6). With the help of this method, crust-mantle interface was determined as 33 km. The calculated depth was verified with the previous seismological (Zhu *et al.* 2006; Akyol *et al.* 2006) and gravity studies (Ankaya & Akçığ 1998). As a result t_c was taken as 33 km.

By benefiting from these approaches and using equation (2) and (8), observed and theoretical admittance curves for various T_e values were calculated (Figure 7). As seen in Figure 7, when theoretical and observed values are evaluated together, the best fit in long wavelengths is for T_e= 6 km. Besides, the Airy compensation model determined with T_e= 0 in Figure 7, is not suitable for the region.

This obtained result was evaluated with the penalty function values that are derived by using equation (9) (Figure 8). Figure 8 shows that the best fit for admittance is obtained for t_c= 33 km and T_e= 6 km because of H_f minima.

Discussion and Conclusion

According to the elastic plate model, the lithosphere is gently flexed into broad upwards and downwards in the region of large loads. The warping induces bending stresses. These stresses will be relieved by brittle faulting in the upper crust and by some form of ductile flow in its lower part. In between the brittle and ductile deformation fields there is an elastic core, which apparently is able to support the stresses induced by flexure on long geological time-scales (Watts 2001; Watts & Burov 2003). Thus, some relationship might exist between the seismogenic layer and T_e. For this purpose, the relationship between the effective elastic thickness and the earthquake focal depths of the Western Anatolia was analysed.

The histograms were formed for the distribution of the focal depths of the earthquakes with M ≤ 4 between the years 1900–2007 that were obtained from Incorporated Research Institutions for Seismology (IRIS)

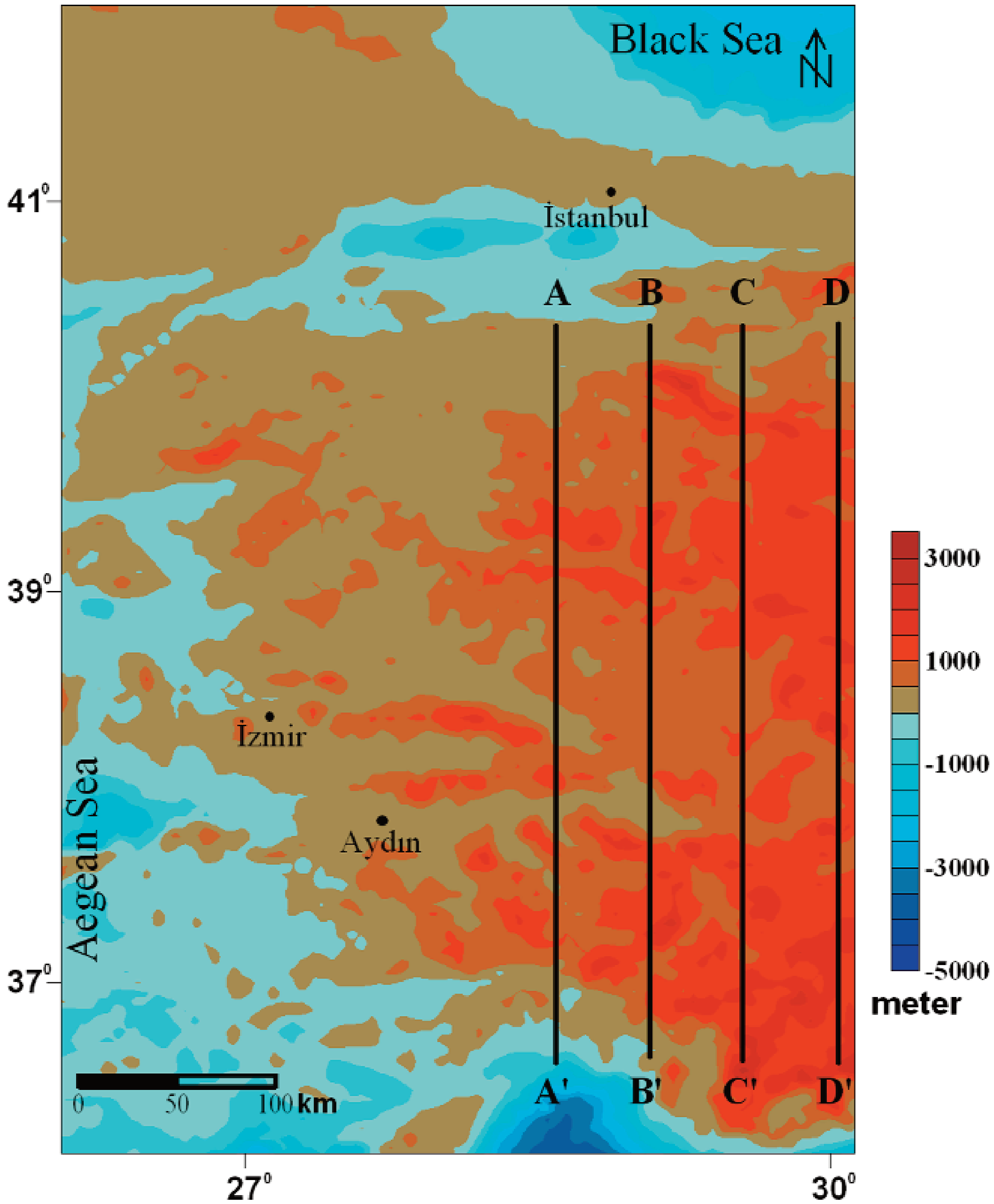

Figure 3. Topographic map of the field and A–A', B–B', C–C', D–D' profiles.

(Figure 9). As seen in Figure 9, the seismogenic zone exists in the first 10 km of the crust. This result is consistent with T_e= 6 km. However, seismogenic layer thickness reflects the strength of, or more precisely the stress level in, the uppermost brittle layer of the lithosphere while T_e is indicative of the strength of the elastic portion of the lithosphere. As a result of this part, thickness of the seismogenic crustal layer of Western Anatolia correlates well with the T_e. The T_e may depend on thermal gradient of the lithosphere. An increase in heat flow in a region can be explained by an increase in thickness of heat producing layers in the crust (Pinet *et al.* 1991). T_e values might decrease in regions where heat flow increases (Hartley *et al.* 1996).

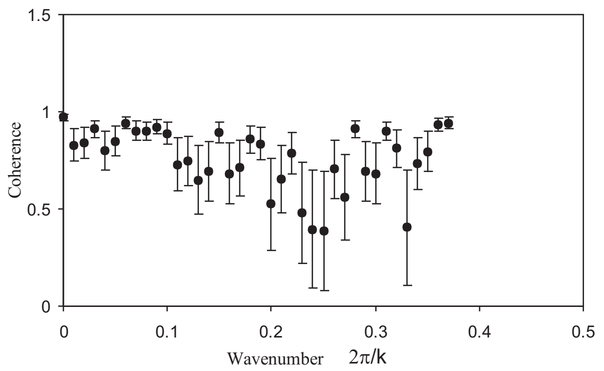

Figure 4. The coherence between the Bouguer gravity anomaly and topography.

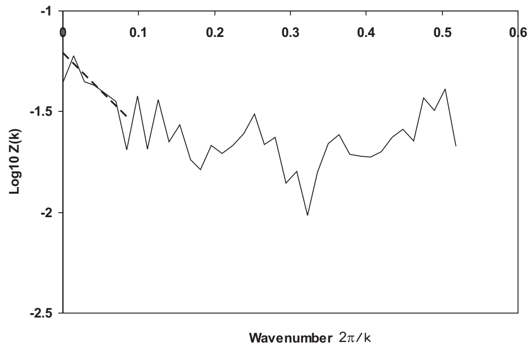

Figure 5. The admittance is determined from spectral analysis from topography and gravity anomalies. Dashed line shows the location of the projected profile data used to obtained smooth spectral estimates.

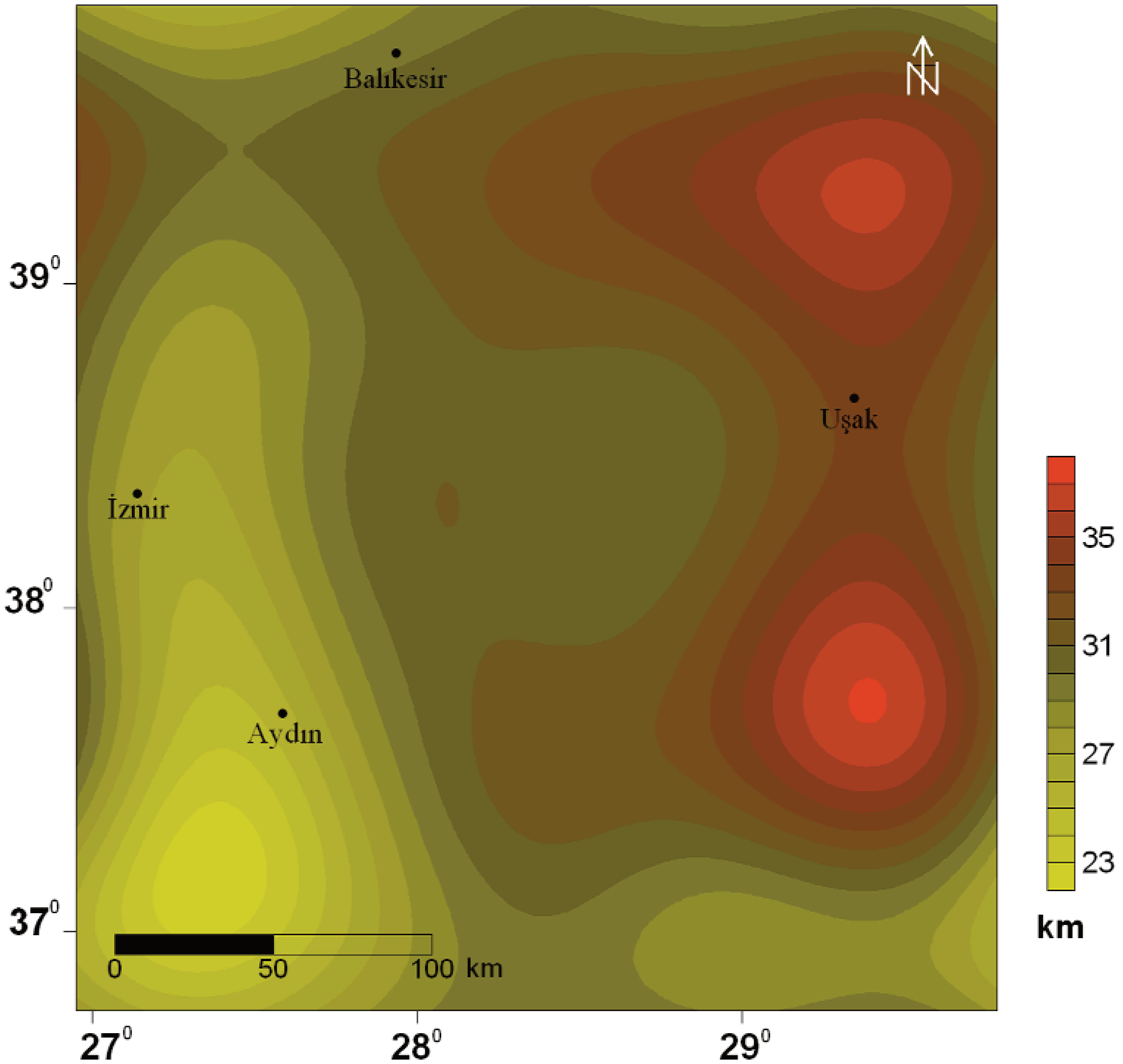

Figure 6. Crust-Mantle interface values.

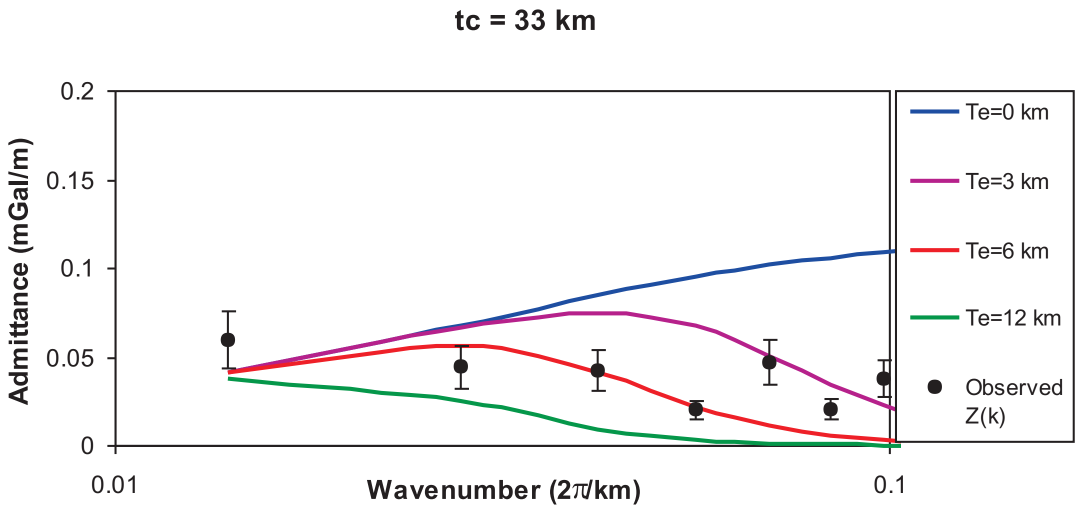

Figure 7. Admittance obtained from equation (2) for Bouguer anomaly and topography.

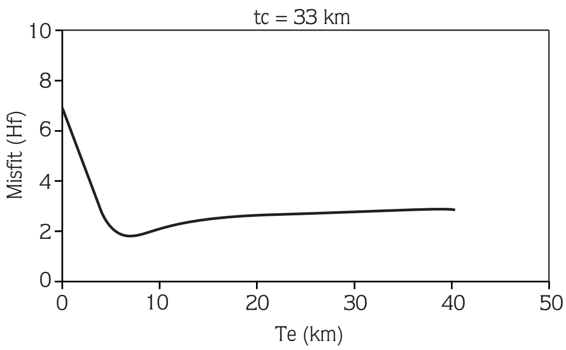

Figure 8. The misfit (H_f) obtained from equation (9) for t_c= 33 km.

According to Dolmaz *et al.* (2005) the thermal anomaly of Western Anatolia is interpreted as a result of asthenospheric upwelling in response to lithospheric extension in this region. This is an indicative of a possible problem from the aspect of T_e and rigidity in the lower crust. When Curie depth and heat flow studies which was done in the region were investigated (Tezcan & Turgay 1989; Dolmaz *et al.* 2005; Şalk *et al.* 2005), especially in the regions where the effective elastic thickness was calculated, thermal depth is shallow and the heat flow is high. Moreover, in a deep seismic study performed by Çifçi *et al.* (2000) for the same region, the strongest seismic reflection comes from the first 6 km averagely.

All these consequences are the isostatic model of the Western Anatolia region which does not fit the local Airy model and are consistent with the finding that 6 km of the Western Anatolian lithosphere may be more resistant to the stresses induced by long time scaled geological flexure.

Besides these, in the region that corresponds to high topography in Figure 3 and low amplitude Bouguer gravity anomaly in Figure 2, there is no significant increase in the depth of crust-mantle interface (Figure 6). In addition, the effective elastic thickness in the region is lower than the world average of similar area (Maggi *et al.* 2000; Watts 2001). As a result, when all these findings are evaluated together with extending tectonics in the region, deformation may affect the strength in the crust on long time scale.

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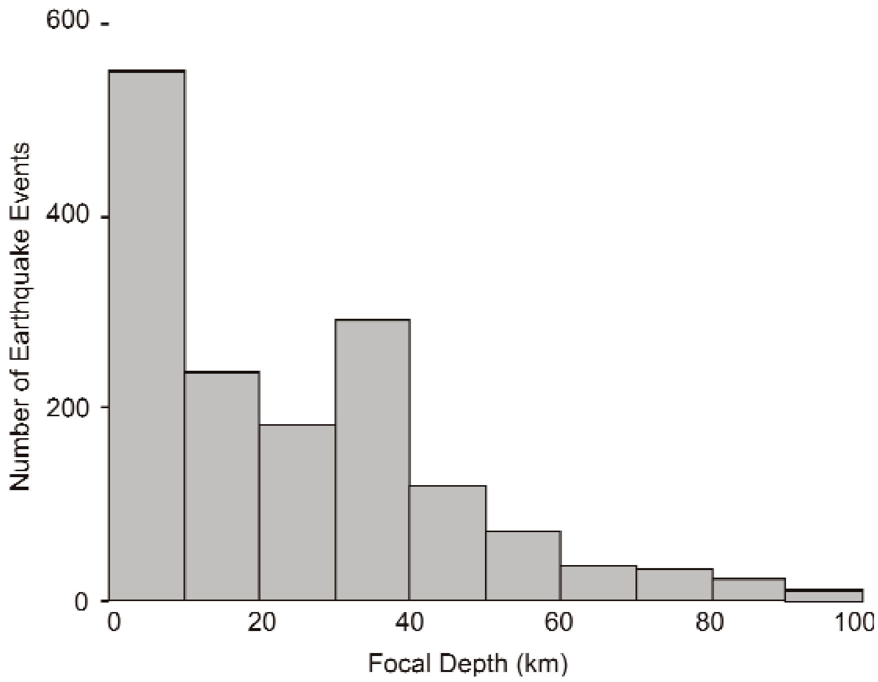

Figure 9. The histogram shows the distribution of earthquakes with focal depth.

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