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Application of boundary analysis and modeling methods on Bouguer gravity data of the Gediz Graben and surrounding area in Western Anatolia and its tectonic implications

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Abstract: *Western Anatolia has been deformed by disintegrating under the effect of N-S oriented tensile forces since the Miocene. As a result of this deformation, many depressions have developed in NE-SW and E-W directions. These are Gediz and Büyük Menderes Grabens in E-W direction and Gördes, Selendi and Demirci basins towards north orienting in NE-SW direction. The relation of these basins, developing in the different directions, has been widely discussed. In order to determine the extensions, continuities and depth variations of these intersecting structures, boundary analysis and basement topography methods were applied to the Bouguer gravity data. By using obtained parameters, graben structures were modeled according to the Talwani method. As a result of basement topography and modeling studies, average thickness of the basin fill units varies from about 2 km to about 3 km in the Gediz Graben, and 3-4 km in the middle parts. According to the boundary analysis method, the Gördes, Demirci and Selendi basins at the north continue within the Gediz Graben towards the south. These findings also prove the existence of NE-SW oriented grabens developed in Early Miocene, within the E-W oriented Gediz Graben.*

Key words: *Gediz Graben, gravity, boundary analysis, basement topography*

INTRODUCTION

Western Turkey is one of the most seismically active regions of the world where orogenic contraction is followed by continental extension (Dewey and Şengör, 1979; Jackson and McKenzie, 1984; Şengör et al., 1985; Eyidoğan and Jackson, 1985; Şengör, 1987; Seyitoğlu and Scott, 1991; Bozkurt, 2001). Two types of basins trending at E-W and NE-SW direction are present in western Turkey (Şengör et al., 1985; Yılmaz et al., 2000). E-W-trending Plio-Quaternary basins are characterized by lateral alluvial fan and axial river deposits that fill the graben floor. The best studied basins belong to the Gediz graben and are bounded by seismically active high-angle normal faults (Arpat and Bingöl, 1969; Şengör et al., 1985; Eyidoğan and Jackson, 1985). Their activity is shown by numerous earthquakes. In the north of the Gediz graben, there are the NE-trending basins. The best known of these are Gördes, Demirci, Selendi, and Uşak-Güre basins (Fig. 1).

The region is extended in N-S direction since at least the Early Miocene. The N-S extension in western Anatolia is attributed to tectonic escape-related deformation which caused westward motion of the Anatolian plate along its boundary structures; dextral North Anatolian Fault Zone and sinistral East Anatolian Fault Zone (Şengör et al., 1985; Seyitoğlu and Scott, 1992). E-W trending normal faults bounded by the hanging-wall blocks appear to root into (Seyitoğlu et al., 2002) or cut the presently low-angle normal faults (Koçyiğit et al., 1999; Yılmaz et al., 2000) which are commonly exposed in culminations known as Menderes Metamorphic core complex (Bozkurt and Park, 1994).

The extension in western Anatolia has been attributed to several tectonic causes, including (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 and 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 and Scott, 1992), 'two stage basin formation' model: 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 (Koçyiğit et al., 1999; Bozkurt and Sözbilir, 2004; Bozkurt and Sözbilir, 2006).

Geophysical studies are not sufficient although many geological investigations were performed in the region. Akçığ (1983) determined an average depth of 2.5 km for the Gediz Graben by using power spectrum and inverse solution techniques to the Bouguer gravity data. A similar study was conducted by Sarı and Şalk (2006) by using the same algorithm techniques. Gürer et al., (2002) indicated that the Early Miocene-aged grabens intersect with the E-W striking Gediz Graben.

At the initial step of this study, the boundary analysis method, developed by Cordell and Grauch (1985) and modified later by Blakely and Simpson (1986), was applied to investigate the Early Miocene aged NE-SW oriented grabens. Then, depth variations in grabens were determined by using basement topography method proposed by Murthy and Rao (1989) and Murthy et al., (1990). In the final stage, benefiting from the parameters obtained, the geological models of the studied area were interpreted based on the Talwani method (Talwani et al., 1959).

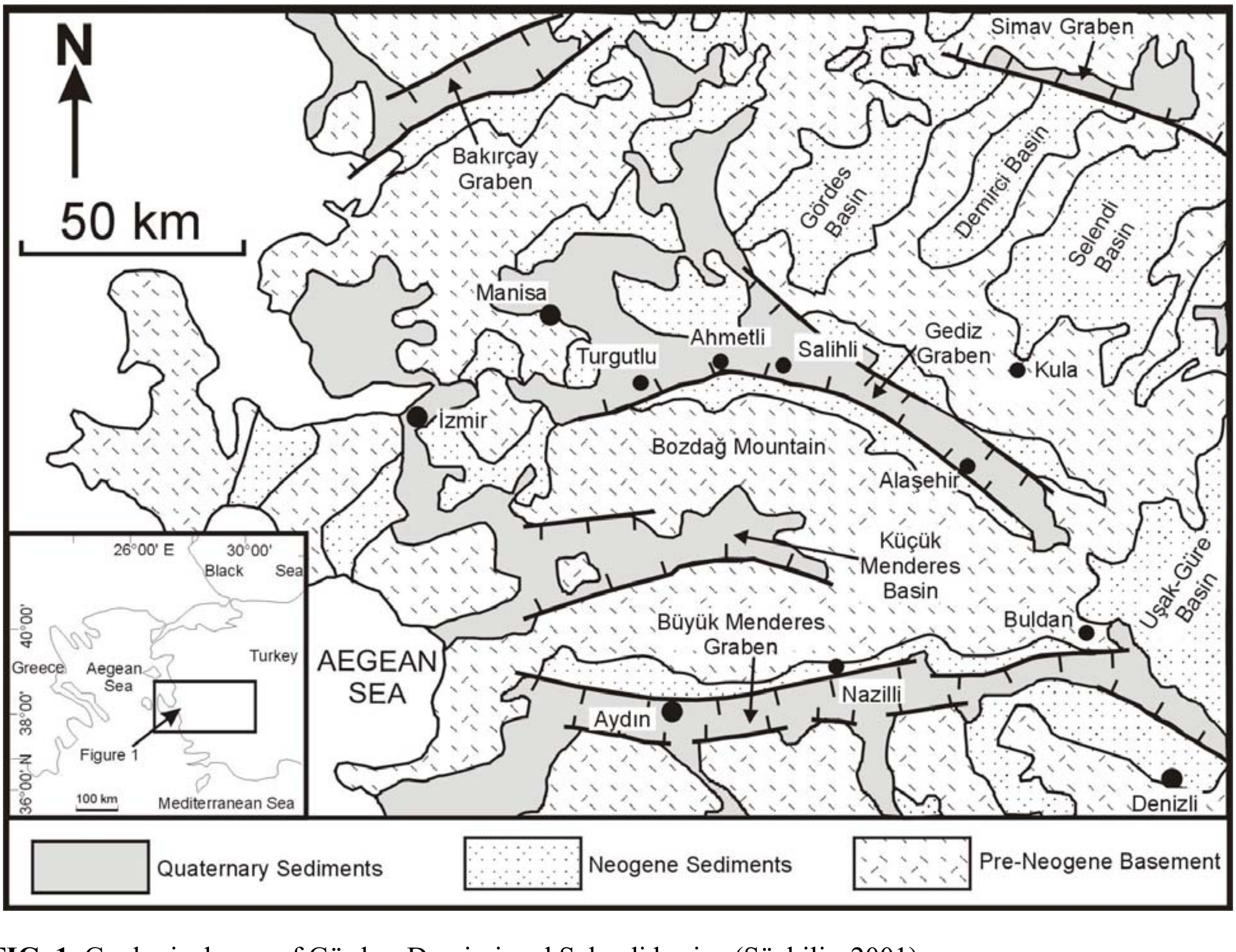

FIG. 1. Geological map of Gördes, Demirci and Selendi basins (Sözbilir, 2001).

THE GEDİZ GRABEN-GEOLOGY

The published studies concerning the geology of the western Anatolian grabens have increasingly continued since Arpat and Bingöl (1969). Most of these studies have focused on the E-W trending Gediz graben because of the following features.

The Gediz graben which is also known as the Alaşehir graben form is 150-km-long, 3-40-km-wide, has an approximately E-W-trending structure and forms one of the most prominent arc shaped, structural elements of western Turkey (Bozkurt and Sözbilir, 2004; Bozkurt and Sözbilir, 2006). It is an actively growing asymmetric graben, being steeper along the southern side of the graben where Bozdağ Mountain rises to over 2000 m from the present-day graben floor, which is 100 m above sea level. With the active normal faults mainly located on the southern margin (e.g., Alaşehir earthquake of 28 March 1969: Arpat and Bingöl, 1969; Eyidoğan and Jackson, 1985), there are reports of relatively less important segmented normal faults in the northern margin (e.g., Roberts, 1988; Paton, 1992; Yusufoğlu, 1996). The southern margin of the basin was uplifted formerly along the presently low angle normal faults, but later the margin was elevated by the high-angle normal faults which cut and exposed the former one (Koçyiğit et al., 1999; Sözbilir, 2001; Sözbilir et al., 2003; Bozkurt and Sözbilir, 2004).

The extension along the detachment fault in the Gediz graben began at 19.5±1.4 Ma (Hetzel et al., 1995a; Hetzel et al., 1995b) and continued until 7±1 Ma (Lips et al., 2001). The extension direction is NNE, based on mylonitic lineation (Hetzel et al., 1995a; Hetzel et al., 1995b; Işık et al., 2003), fault slicken lines (Hetzel et al., 1995a; Hetzel et al., 1995b; Koçyiğit et al., 1999; Sözbilir, 2001; Sözbilir, 2002; Bozkurt and Sözbilir, 2004), and the fold axis parallel to extension is oriented in NNE direction (Sözbilir, 2001).

In this area, the E-W-trending low-angle normal fault poses a distinctive sequence of low- and high-grade metamorphics, Tethyan ophiolites and Miocene sediments against a footwall of metamorphic rocks of the Menderes Massif. The Gediz fault is a crustal-scale, N-NE-dipping and presently low-angle breakaway fault that forms the structural boundary, along the southern Gediz graben, between synextensional sediments in the hanging wall and the crystalline basement in the footwall (Hetzel et al., 1995a; Hetzel et al., 1995b; Emre, 1996; Koçyiğit et al., 1999; Lips et al., 2001; Sözbilir, 2001; Sözbilir, 2002; Bozkurt and Sözbilir, 2004). The Gediz detachment which can be traced discontinuously along the strike for more than 100 km from Turgutlu to Alaşehir district is one of a crustal-scale detachment faults that formed during the Miocene in the western Anatolian extensional province. This fault has a gently undulating shape, reflected by the geomorphology. The undulations in the detachment fault have been interpreted as original corrugations in the fault surface, because they are parallel both to the inferred movement direction of the upper plate and the extension lineation in lower plate mylonites (Koçyiğit et al., 1999; Sözbilir, 2001). The crystalline basement in the Bozdağ Mountain includes metamorphic rocks of predominantly Palaeozoic age that have been intruded by Miocene granites (Hetzel et al., 1995a; Hetzel et al., 1995b; Hetzel et al., 1998; Koçyiğit et al., 1999; Yılmaz et al., 2000; Sözbilir, 2001; Okay, 2001). The structural top of this formation is marked by the Gediz fault which characteristically defines dome-and-basin structural topography (Sözbilir, 2001). This fault was named by Emre (1996) and Koçyiğit et al., (1999) as the "Karadut detachment fault" and the "Çamköy detachment fault", respectively. The term "Gediz detachment fault" was introduced by Lips et al., (2001) to refer to the master detachment under which the Menderes metamorphic core complex was exhumed (see Bozkurt and Oberhansli, 2001 for detail discussion of the Menderes Massif).

The presently exposed detachment faults in the southern margin of the Gediz graben are truncated and displaced by high-angle normal faults indicating "rift mode" extension in western Turkey (Koçyiğit et al., 1999; Sözbilir, 2001; Bozkurt, 2001; Bozkurt and Sözbilir, 2004).

The stratigraphic units of the study area have been described by Cohen et al., (1995), Dart et al., (1995), Emre (1996), Seyitoğlu and Scott (1996), Koçyiğit et al., (1999), Sarıca (2000), Seyitoğlu et al., (2000; 2002). Yılmaz et al., (2000) subsequently classified a stratigraphic scheme that consists of sequences I through IV (Sözbilir, 2001; Sözbilir, 2002; Bozkurt and Sözbilir, 2004). All sequences are bounded by angular unconformities. In brief, sequence I consists of alluvial fan delta deposits that rest unconformably on pre-Tertiary upper-plate rocks. Sequence II consists of reddish alluvial fan-braided river deposits which have been dated at late Miocene-Early Pliocene based on palynological and palaeontological methods. These deposits accumulated during supradetachment basin formation produced by a NNE-dipping low-angle detachment fault. A depositional model for late sedimentation is that of alluvial fans deposited into E-W-trending basin, and deposition of gravel-sand in axial river that flowed to the West along the axis of the basin. Sequence III mostly contains semi-lithified conglomerate-sandstone which has been dated at Pleistocene based on mammalian fauna. Sequence IV consists of lateral alluvial fans and axial river deposits that fill the present graben floor. To the south of Ahmetli (Turgutlu) area, Sözbilir (2002) mapped several exposures of metamorphic rocks in the upper-plate of the Gediz detachment fault and interpreted them as slices of the metamorphic extensional allochthons which consist of various schists with lenses of amphibolites and blocks of rudist-bearing marbles. In the southeast of Salihli, around Dağahmetli village, two orthogneiss slices rest on the Gediz detachment fault. These gneisses were defined as a klippe by Hetzel et al., (1995a; 1995b). However, field observations and kinematic studies (Emre and Sözbilir, 1997; Lips et al., 2001) support the opposite idea that the klippes belong to one of the upper-plate slices of the detachment fault. North of the Gediz graben, Early-Middle Miocene volcano-sedimentary successions are exposed along the NE-trending depressions (Seyitoğlu, 1997; Bozkurt, 2003). These depressions recently have been explained as km-scale corrugations of the Gediz detachment fault (Purvis and Robertson, 2004; 2005a; 2005b).

METHODS

Boundary analysis method

The method was first used by Cordell and Grauch (1985) to determine the density of rocks, locations of sudden changes in the magnetization of rocks and the boundaries of these source structures. Then, it was developed by Blakely and Simpson (1986) and automated so as to determine the locations of horizontal gradient amplitudes over the contoured map plane.

First of all, horizontal gradient amplitudes are found from the Bouguer gravity anomaly. If g*(x,y)* is the gravity field then the horizontal gradient magnitude *HGM(x,y)* is given by:

$$HGM(x,y) = \sqrt{\left(\frac{\partial g(x,y)}{\partial x}\right)^2 + \left(\frac{\partial g(x,y)}{\partial y}\right)^2} \quad (1)$$

where $\partial g(x,y)/\partial x$ and $\partial g(x,y)/\partial y$ are the two first-order horizontal derivatives of the gravity field, respectively. Later on, by using the adjacent 8 spots given in Figure 2, the maximum value and the location of the horizontal gradient is computed by:

$$x_{\max} = \frac{-bd}{2a} \quad (2)$$

where d is the grid interval and parameters a- and b- are defined as:

$$a = \frac{1}{2}\left(HGM_{i-1,j} - 2HGM_{i,j} + HGM_{i+1,j}\right) \quad (3)$$

$$b = \frac{1}{2}\left(HGM_{i+1,j} - HGM_{i-1,j}\right) \quad (4)$$

The horizontal gradient at $x_{\max}$ point is:

$$HGM_{\max} = ax_{\max}^2 + bx_{\max} + HGM_{i,j} \quad (5)$$

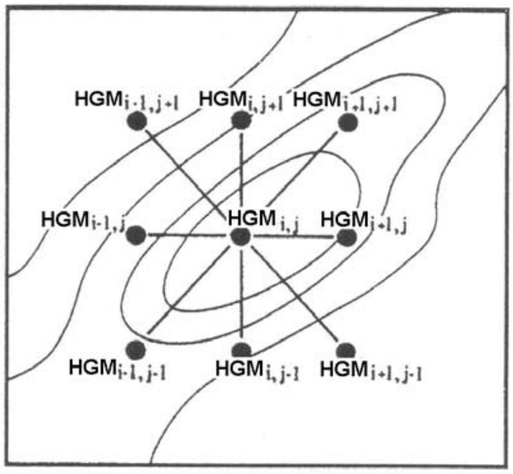

FIG. 2. Location of the spots on the map (Blakely and Simpson, 1986).

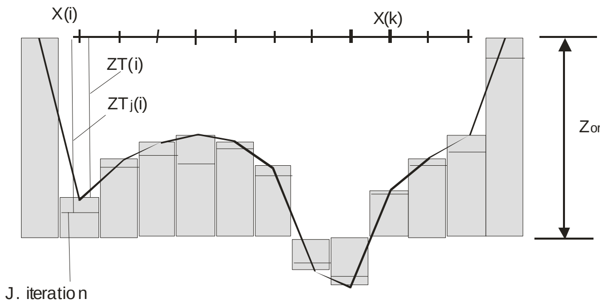

FIG. 3. Prisms representing the basement topography.

Determination of Underground Basement Topography by the Inverse Solution of Gravity Anomalies

The depth to the upper surface of the prisms, in which a structural model is defined by adjacent prisms extending to the infinite in the vertical direction, can be found by using Δg Bouguer gravity anomalies that are sampled with equal intervals (Fig. 3).

The relation $\Delta f(k) = \Delta f(X_k)$ giving the gravity anomaly of a dike at any point of surface $P(X_k)$ is defined as (Rao and Murthy, 1978; Murthy et al., 1990):

$$\Delta f(k) = \sum_{i=2}^{N-1} \left[F_k(z)\right]_{ZT(i)}^{Z_{ave}} + AX_k + B \quad (6)$$

where *ZT(i)* is the depth beneath the i^th^ sampling point and Z_{ave} represents the average depth of underground basement topography along the profile. The function $F_k(z)$ is defined as follows in the gravity method (Rao and Murthy, 1978):

$$F_k(z) = 2G\sigma\left[Z\left[\tan^{-1}\left(\frac{X_k + \frac{dx}{2}}{Z}\right) - \tan^{-1}\left(\frac{X_k - \frac{dx}{2}}{Z}\right)\right]\right.$$

$$+0.5\left[\left(X_k + \frac{dx}{2}\right)\ln\left(\left(X_k + \frac{dx}{2}\right)^2 + Z^2\right)\right.$$

$$\left.\left.-\left(X_k - \frac{dx}{2}\right)\ln\left(\left(X_k - \frac{dx}{2}\right)^2 + Z^2\right)\right]\right] \qquad (7)$$

Here, G is the gravity constant, σ denotes the density contrast and dx represents the sampling interval. The method is applied in two stages as the initiation stage and the second stage, in which the results are adjusted by repetition. $ZT_j(i)$ (i = 2,3, …, N) should be the initial depths of the prisms to the surface. If A_j and B_j are the coefficients of regional anomaly, then the theoretical anomaly of the model at the end of j times repetition can be written as:

$$\Delta f_j(k) = \sum_{i=2}^{N-1}\left[F_k(z)\right]_{ZT_j(i)}^{Z_{ave}} + A_j X_k + B_j \qquad (8)$$

In the construction of the repetition plane, the depth of the upper surface of each prism is initially taken as the average depth. The differences of prism depths from the average depth are $\left[Z_{ave} - ZT_1(i)\right]$ and rate of change of anomaly is $\partial F_k(z)/\partial z$. According to Equation (8)

$$\Delta f_k(k) = \sum\left[\frac{\partial F_k(z)}{\partial z}\right]_Z \left[Z_{ave} - ZT_1(i)\right] + A_1 X_k + B_1 \qquad (9)$$

Initial values of the depth of the prisms from the surface $ZT_1(i)$ and the regional anomaly coefficients A_1, B_1 can be calculated from Equation (9). By using Equations (6) and (9), the difference between the observed and computed anomaly at the end of j repetitions at point (X_k) can be given below:

$$\mathrm{df(k)} = \sum_{i=2}^{N-1}\left[\mathrm{F_k(z)}\right]_{\mathrm{ZT_j(i)}}^{\mathrm{ZT(i)}} + \mathrm{dAX_k} + \mathrm{dB} \qquad (10)$$

In this equation; dA and dB symbolize the increments to be given to the regional anomaly coefficients. In case the amount dZT(i)=ZT(i)-ZT_j(i) is insignificant, Equation (10) becomes:

$$\mathrm{df(k)} = \sum_{i=2}^{N-1}\left[\frac{\partial \mathrm{F_k(z)}}{\partial z}\right]_{\mathrm{ZT_j(i)}} \mathrm{dZT(i)} + \mathrm{dAX_k} + \mathrm{dB} \qquad (11)$$

Hence, a convenient optimization method can be applied to the above equations by minimizing the $\sum(\mathrm{df(k)})^2$ error function and solving for every anomaly point (X_k).

APPLICATIONS

In this study, we utilized Bouguer gravity values (MTA, 1979) from a region which covers the Gediz Graben and related basins, and located between 38°-40° N latitudes and 27°-29° E longitudes. First degree trend was applied to the numerical data with a sampling interval of 1 km to eliminate the influence of basic regional trend. The deduced map is depicted in Figure 4.

As the first step, the boundary analysis method was applied to determine the tectonic borderlines in the region. From the horizontal gradient values (in x and y directions), maximum horizontal gradient values were calculated. The obtained maxima are plotted on the map (Fig. 5 and 6).

The sequence of maximum gradient values shows great coherence (Fig. 5, 6) with the structural borders (Gediz, Gördes, Demirci, Selendi and Simav Grabens) observed in the north of the study area.

In the second step, the basement topography method was applied to the same data and the depth variations of the region were determined. In practice, the densities in the region are selected as 2.2 g/cm^3 for the sediments and 2.7 g/cm^3 for the Paleozoic Menderes metamorphics forming the basement rock. Several example profiles relevant to the obtained results are given in Figure 7 whereas cumulative outcomes are illustrated in Figure 8.

The depth values happen to be about 4 km in the middle part of Gediz Graben (Fig. 7 and 8) and occasionally reach 2-2.5 km in the northern and southern parts of the graben. This outcome is also consistent with the results of a recent study carried out by Gürer et al., (2002). The determined depths are indicative of the fact that NE-SW striking basins (Gördes, Demirci and Selendi) in the north of Gediz Graben seem to extend in the southern part of the graben also.

At the final stage of our study, with the findings that were obtained from boundary analysis and basement topography applications, Talwani modeling was accomplished at totally 8 profiles. Some of the results obtained from those applications are given in Figure 9.

The results deduced from this application overlap with the other existing data. When all the findings are evaluated together, the results are in good correlation with the structural model suggested by Yılmaz et al (2000) as in Figure 10.

While the boundary analysis method was used to determine the structure boundary on horizontal direction, it is not used to define the structure boundary on vertical direction. Density information of the media is not important in this method. As known, Talwani modeling and basement topography method are being used to reveal the depth of the structure. However, the density selection is important in the Talwani modeling and basement topography procedure. This parameter affects directly the depth structure. In the present study, the boundary analysis method was used to determine the location of the structure boundaries and faults, the Talwani method was chosen to determine the basin topography.

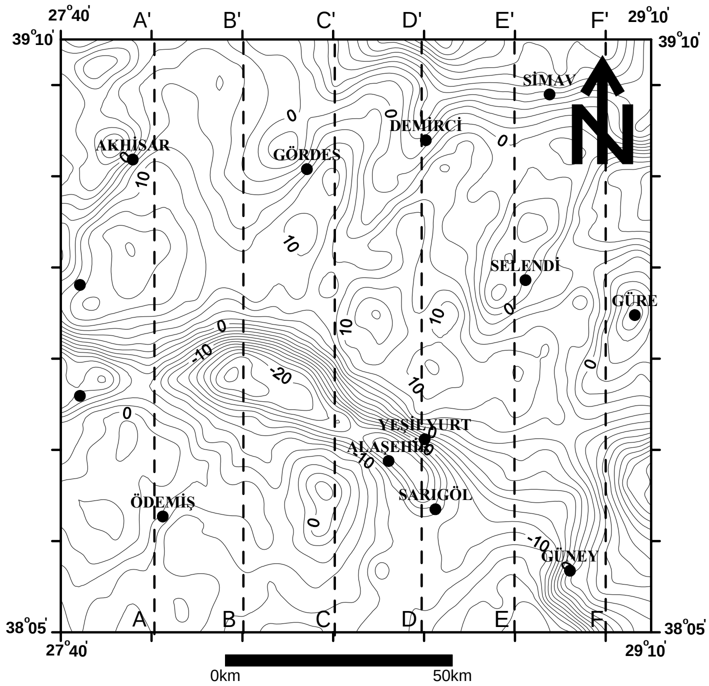

FIG. 4. Bouguer gravity map of the 1st degree trend.

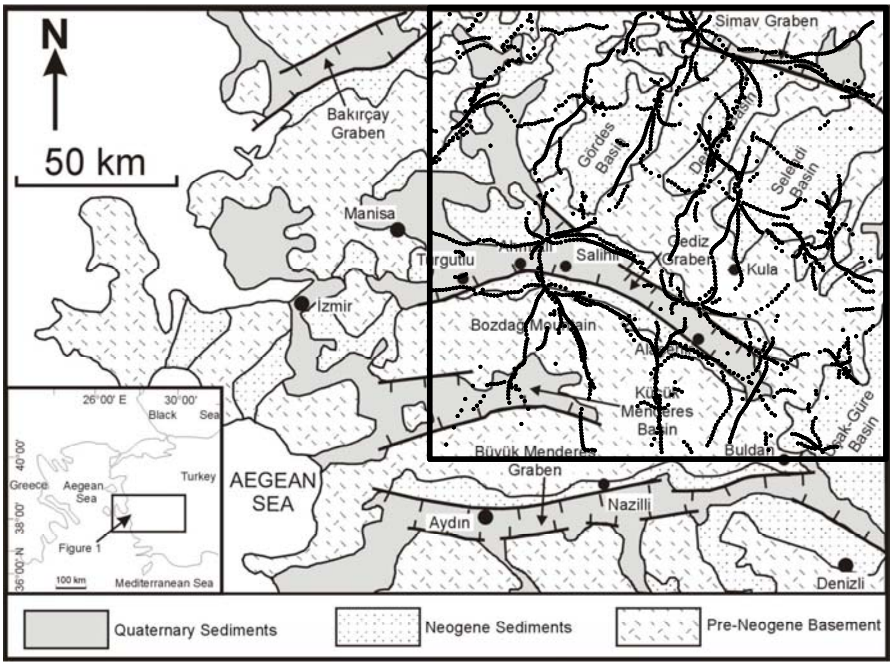

FIG. 5. Demonstration of boundary analysis on the geological map.

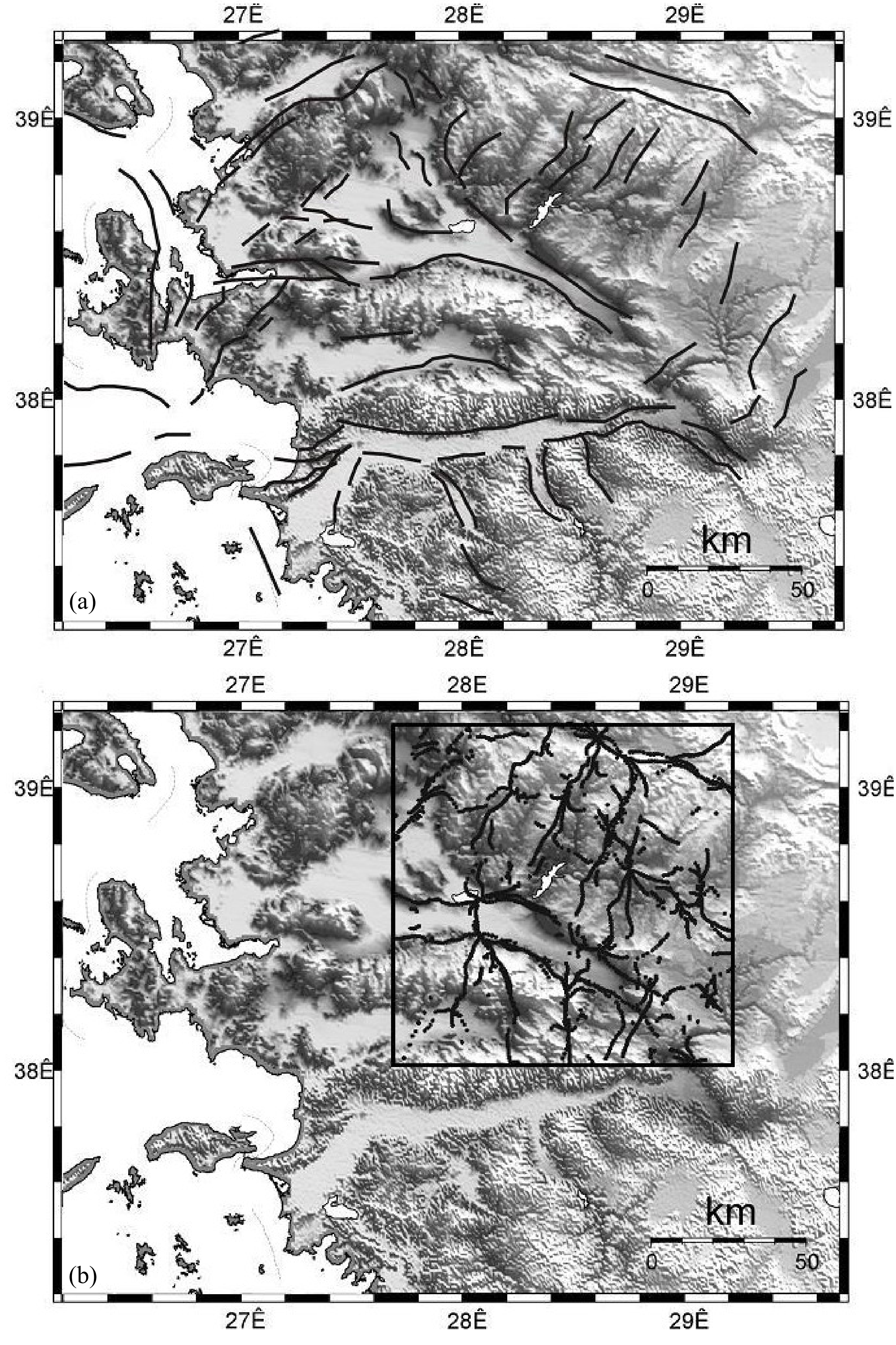

FIG. 6. a) Major structural elements of Western Anatolia (modified from Bozkurt, 2001) b) Demonstration of boundary analysis on the major structural elements of Western Anatolia.

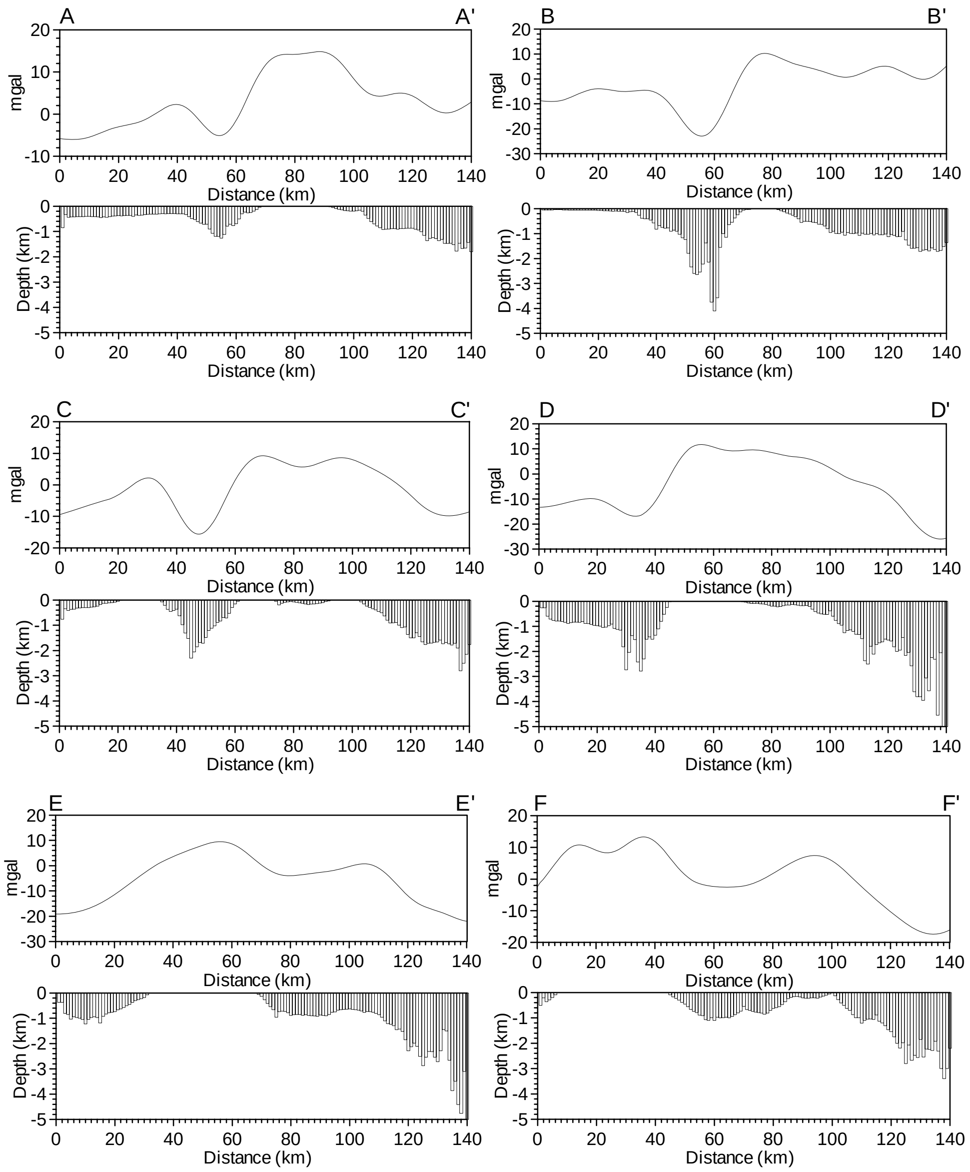

FIG. 7. 1D basement depth variations.

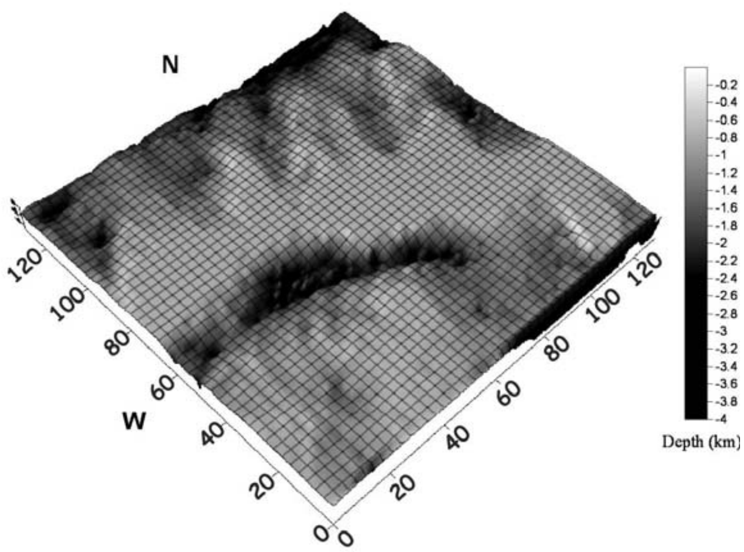

FIG. 8. Cumulative results obtained by the basement topography method.

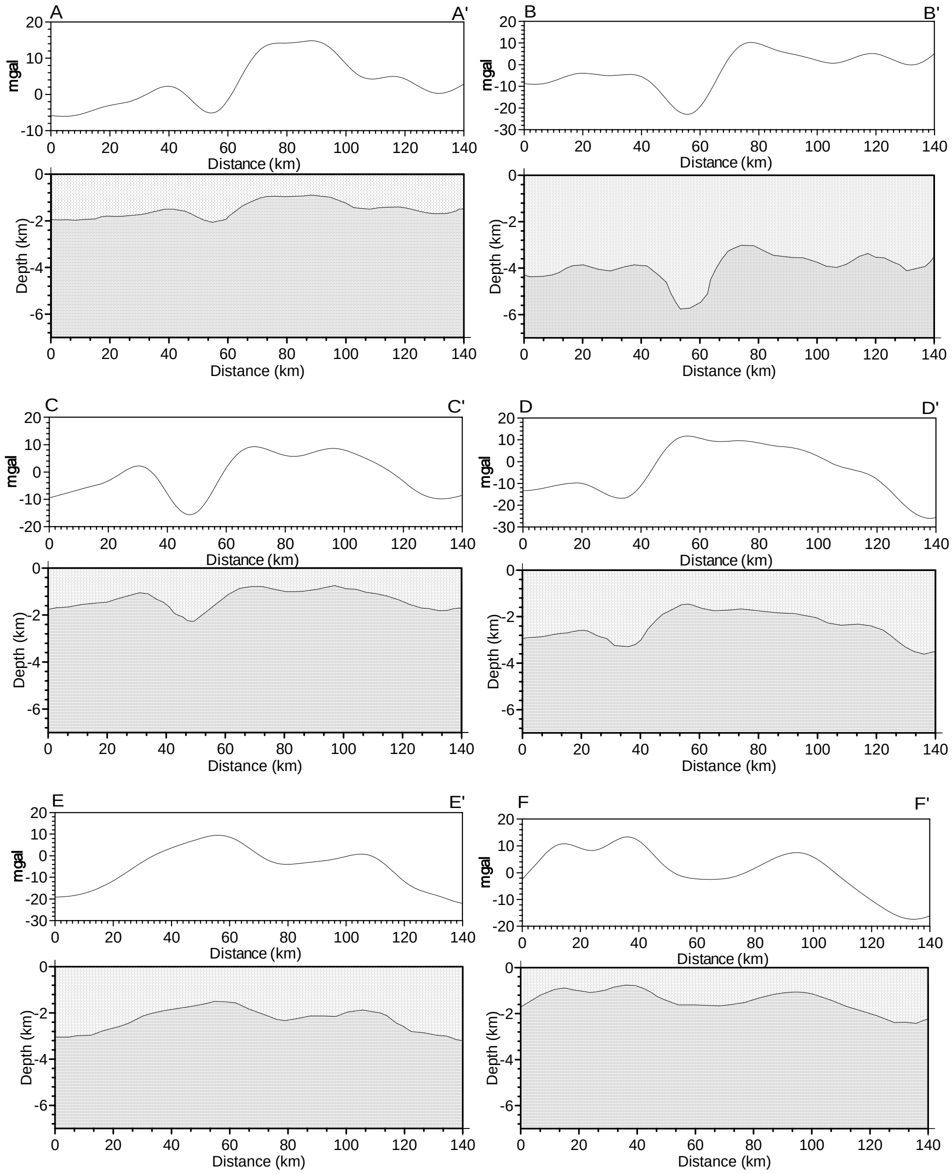

FIG. 9. Outcomes of Talwani modeling.

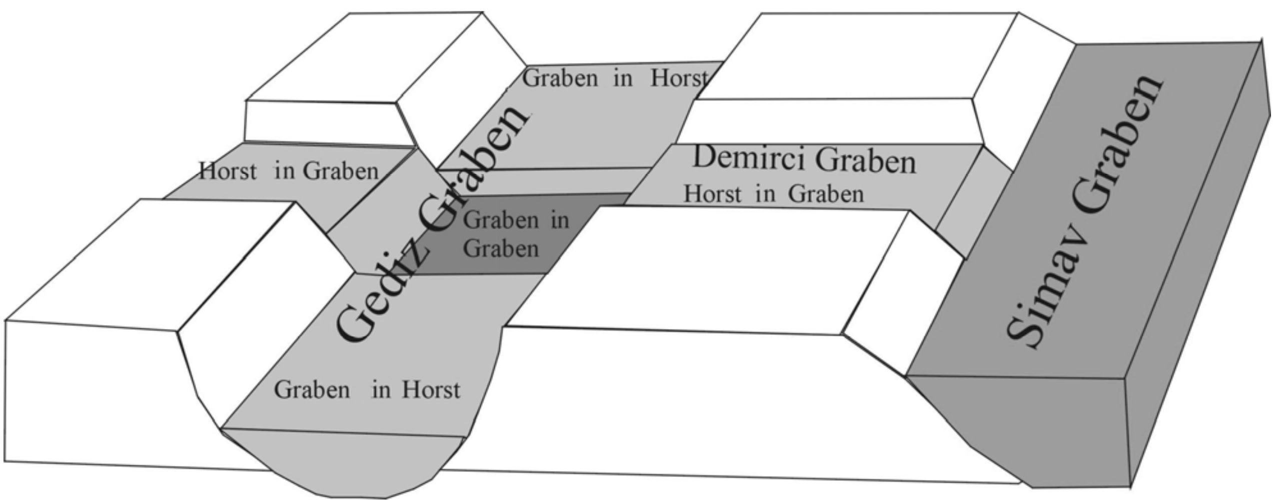

FIG. 10. Suggested structural model for the region (Yılmaz et. al., 2000).

RESULTS AND DISCUSSION

The origin of the NE-trending basins has been also a subject of controversy for many years and is attributed to one of four different models (1) the basins are "palaeotectonic Tibet type cross-grabens" developed under N-S palaeogene compression and shortening across the Neotethys. They are later replaced by Neotectonic Aegean-type cross-grabens (replacement structures) under the following N-S extensional tectonic regime commencing in the Tortonian (Şengör et al., 1985; Şengör, 1987). (2) The second model suggests that these two different-trending basins developed coevaly during a Neotectonic N-S extensional regime, since the latest Oligocene-Early Miocene times (Seyitoğlu and Scott, 1991). The extension is attributed to the late orogenic extensional collapse of the Menderes Massif. (3) The third model considers that NE-trending basins are not related with extension, but are intermontane basins that developed on the collision-related pre-Mesozoic basement (İnci, 2002). (4) Recent view (Purvis and Robertson, 2004; 2005a; 2005b) claims that the depression, in which the NE-trending basin has formed, is related to the primary corrugation of the Gediz detachment fault. According to this model, The footwall of the Gediz detachment is corrugated (wavelength ca. 30 km; amplitude ca. 1.5 km) giving rise to N-S trending structural highs separated by depressions that later became depocenters for the Gördes, Demirci and Selendi basin fills.

Under the light of the above models, the sedimentary coverage of Western Anatolia was determined from the basement topography and boundary analysis method of the Bouguer gravity values (MTA, 1979). The results obtained herein are highly consistent with the extension of Gediz, Gördes, Demirci, Selendi and Simav Grabens. According to the model inferred from NE-SW oriented boundary analysis data, these grabens continue towards South starting from the north. They overlap with the forms of principal structures such as Gördes, Demirci, Selendi Grabens in the north. Nevertheless, their continuity towards the south can not be observed on the tectonic maps. This can be explained by the existence of significant deep structures and lack of exposure.

According to the basement topography method, the thickness of the sedimentary deposits varies from about 2 km to about 2.5 km in the north and south of the Gediz Graben. Maximum thickness of the sedimentary column is 4 km occasionally in the mid-part of the graben. This interpretation is also consistent with the results provided by Gürer et al., (2002).

The depths determined in the north of Gediz Graben and overlapping with Gördes, Demirci ve Selendi Grabens reaching up to 4 km in thickness, are not coherent with the structural shapes observed in the south. These depths also symbolize the structural shapes that are not exposed. The models presented here exhibit great coherence with the Talwani method.

All the consequences indicate that Gördes, Demirci ve Selendi Grabens encountered in the north of Gediz Graben also appear to continue in the south of the graben. This conclusion confirms that NE-SW striking grabens in the region are cut by Gediz Graben as proposed by Yılmaz et al., (2000).

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