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Numerical modeling of the conductive heat transfer in western Anatolia

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Abstract: *Using the available data, the crustal temperatures were modeled along two profiles in the western Anatolia. Modeling was achieved by solving a two-dimensional steady-state heat conduction equation in a heterogeneous medium by successive overrelaxation method of finite differences. The Bouguer gravity data were used to construct the crustal structures along the profiles. The average crustal thickness was taken as 30 km. An exponential decrease with depth in heat production was assumed. The temperatures in the grabens and regions under them are obtained as relatively higher than those in the surrounding regions. The thickness of the seismogenic zone and depth to the Curie isotherm in the region were estimated as 10 and 15 km respectively. Also, the temperature at the bottom of the crust was calculated as 1075-1100ºC approximately. Sedimentary fills in the grabens affect the temperature distribution in the upper crust.*

INTRODUCTION

Temperature distribution within the earth's crust has important effects on the active tectonics and seismicity. Even though it is one of the most important parameters in the earth, our knowledge of the thermal state within the crust is poor. Temperature measurements are usually restricted at shallow depths and are under the influence of fluid motions, topography and local geology. Therefore, some mathematical models are widely used to estimate temperature change with depth (Mayhew, 1982). Based on some assumptions, numerical modeling of steady-state conductive heat transfer is one of the methods, which is widely preferred by researchers, to define temperature distribution both horizontally and vertically in the crust and upper mantle (Sams and Thomas-Betts, 1988; Clauser and Villinger, 1990; Baumann and Rybach, 1991; Swift, 1991; Cermak et al., 1991; Kukkonen, and Joeleht, 1996; Jokinen and Kukkonen, 1999). Defining temperature distribution may reveal the characteristics of thermal regime in the crust. Since the thermal regime in western Anatolia is anomalous (Hurtig et al., 1992; IlkiS ik, 1995; Tezcan, 1995), we have carried out a two-dimensional modeling in the region using available data to define temperature distributions along several cross-sections.

The main driving force of active tectonics in Anatolia is the lithospheric regime connected with the continental collision in the Alpine-Himalayan zone. Interaction among the Eurasian, African and Arabian plates controls the active tectonics in the western Anatolia. Until the beginning of the late Miocene the region was under a north-south shortening, and this yielded a thick crust (> 50 km) (Le Pichon and Angellier, 1981; Sengor, 1982; Jackson and McKenzie, 1988; Yilmaz, 1989). Since the late middle Miocene the region has experienced a north-south extension. This caused the thick and partially melted crust (lower parts only) to be stretched. Thus, this generated a thin and brittle crust (~30 km) (Ezen, 1991; Saunders et al., 1998). There are several types of models for the extension mechanism in the region. These are tectonic escape (Dewey and Sengör, 1979; Sengör 1979, 1980, 1987; Sengör et al., 1985), back-arc spreading (Le Pichon and

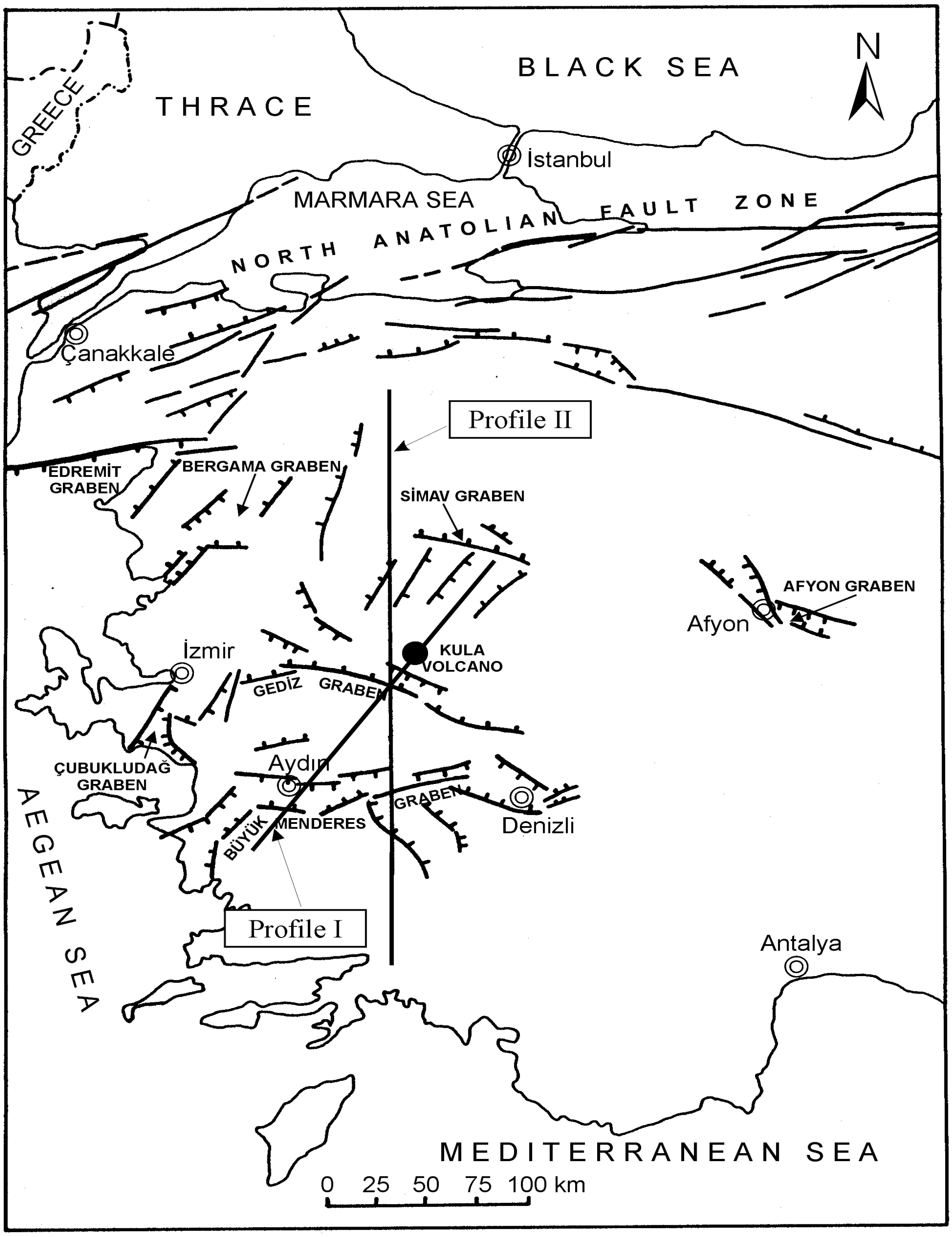

FIG. 1. Major tectonic features and the profiles along which temperature fields were modeled (faults are drawn by B. Rojay from the Landsat images) (Modified from SimSek and Güleç, 1994).

Angelier, 1979; Meulenkamp et al., 1988) and orogenic collapse (Dewey, 1988; Seyitoglu and Scott, 1996; Seyitoglu et al., 1992).

The present geomorphology is characterized by a series of east-west trending major grabens with northeast-southwest trending secondary (cross-cutting) grabens (Koçyigit, et al., 1999) (Fig. 1). Western Anatolia is one of the most rapidly deformed continental regions in the earth, and the present widely spread seismicity is an indicator of this deformation (McKenzie, 1978; Alptekin et al., 1990). However, there is a close association between the present day deformation and the rate of convergence and subduction along the Hellenic-Cyprus Arc (Royden, 1993a,b).

High heat flow and seismicity, intensive faulting and volcanism are the main characteristics of the region. Tezcan (1995) mentions that the heat flow values over Anatolia are much higher than those over the Meditterranean Sea and Black Sea. A similar result was also reported by IlkiSik (1995). He calculated a mean heat flow value as 107±45 mWm^{-2}, which is approximately 50% higher than the world average. Also, the surface temperatures of hot springs in the western Anatolia are generally higher than those in the eastern Anatolia (Erentöz and Tenek, 1968). Moreover, there is substantial thermal activity in the area revealed by numerous hot springs, fumaroles, hydrothermal alterations and recent mineralizations. Major geothermal fields are usually associated with the horst-graben systems and active-subactive volcanism. For instance, the Kizildere and Pamukkale geothermal fields exist in the Büyük Menderes graben. Especially, major geothermal systems exist where the north-south and east-west trending horst-graben systems intersect each other.

A two-dimensional steady-state conductive heat flow modeling was performed along some profiles in the western Anatolia (Fig. 1) to estimate temperature distribution in the crust. Based on the modeling results, the temperatures in the grabens and regions under them are obtained as relatively higher than those in the surrounding regions. The temperature at the bottom of the crust was calculated to be approximately 1075-1100ºC. Also, the grabens affect the temperature distribution in the upper crust.

HEAT FLOW DATA

Unlike western Anatolia, a large number of heat flow measurements is available in Europe and the surrounding seas. According to these measurements, the eastern Mediterranean Sea has relatively low heat flow values mainly in the southern part. Also, low heat flow is observed in the Black Sea. However, high heat flow values are obtained after correcting original observations for the fast sedimentation (Erickson, 1970). In the Aegean Sea three high heat flow anomalies connected with tectonic zones are observed. The highest heat flow (>120 mWm^{-2}) is related to the subduction zone (Fytikas, 1980).

The overall heat flow values over Turkey are generally much higher than those over the Mediterranean Sea and Black Sea (Tezcan, 1995), although a poor heat flow measurement coverage is available in Turkey (Tezcan, 1995; IlkiSik, 1989). The first heat flow map for Turkey was published by Tezcan (1979) using the measurements in the wells drilled for geothermal energy. This map has been revised by including temperatures measured in 204 oil and coal wells mostly drilled in the south-eastern Turkey and Thrace (Tezcan and Turgay, 1991) (Fig. 2). They calculated heat flow using geothermal gradients in exploration wells and assuming a constant thermal conductivity because no thermal conductivity measurements were performed in the wells. The temperature at the bottom of the hole and the one at 1m below the surface were employed to calculate the geothermal gradient. Since most of the wells have been drilled in sedimentary units, a typical value of 2.1 $Wm^{-1}K^{-1}$ has been assigned as the thermal conductivity in all calculations.

According to the map in Figure 2, the highest heat flow values (>140 mWm^{-2}) are observed in the western Anatolia. It is obvious that there is a close association between the heat flow anomalies and horst-graben systems. Central Anatolia, on the other hand, is characterized by values between 70 and 100 mWm^{-2}. There are two heat flow closures with the values greater than 100 mWm^{-2}. The shapes of the anomalies imply that they are very likely related to the young volcanism (Koçak, 1990). The eastern part of the map (not shown in Fig. 2) does not indicate any anomalies. The values range from 70-100

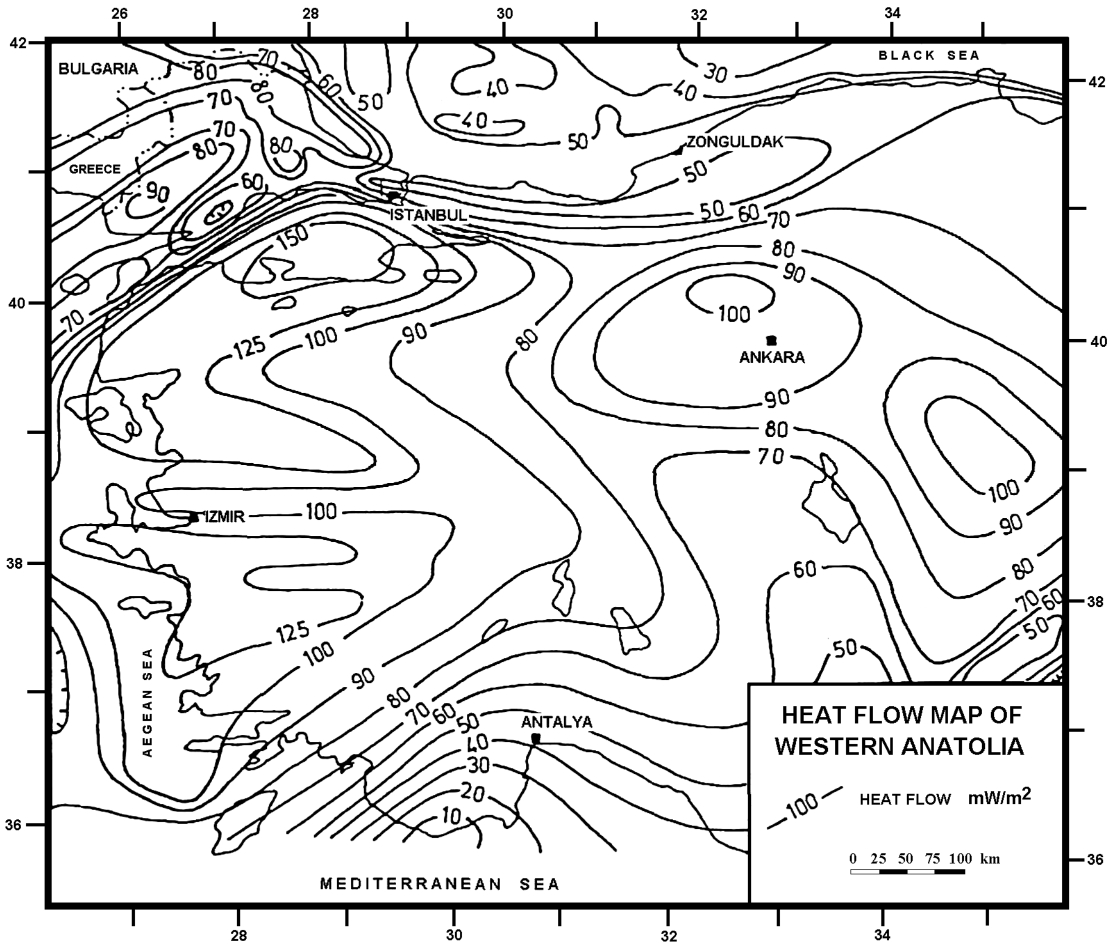

FIG. 2. Heat flow map of western Anatolia (from Tezcan and Turgay, 1991), calculated using geothermal gradients in exploration wells and assuming a constant thermal conductivity ($k = 2.1\ Wm^{-1}\ K^{-1}$)

mWm^{-2}. Since there is no enough sampling in the region, this part of the map may not represent the actual thermal state of the crust in the eastern Anatolia. Due to the available temperature data collected from the wells drilled for oil exploration, the southeastern part of the map (not shown in Fig. 2) is also representative of the thermal regime in the area. The values vary between 50 to 90 mWm^{-2} approximately, and there are some closed contours in the region. Tezcan (1995) considers them to be associated with buried topography.

IlkiSik (1995) determined a regional heat flow pattern in the western Anatolia using silica geothermometers from 187 thermal springs. He calculated the mean value of the heat flow as $107 \pm 45\ mWm^{-2}$. This value is about 50-60% higher than the world average. IlkiSik's (1995) heat flow pattern given in semi-quantitative nature indicates that there is an association between high heat flow and the grabens (the Büyük Menderes, Küçük Menderes and Gediz grabens). Also, there is a close correlation between high heat flow (> 100 mWm^{-2}) and Tertiary and younger volcanism.

Recently, MTA (The Directorate of Mineral Research and Exploration) and TÜBITAK (The Scientific and Technical Research Council of Turkey) together with some universities carried out some heat flow measurements in the western Anatolia using shallow wells drilled for water explorations (which are not in use). This study gives similar results like previous ones. Based on these measurements, an average heat flow is 103.68 mWm^{-2} in the region (IlkiSik et al., 1996).

Following Rybach and Buntebarth (1982), IlkiSik (1995) evaluated an average heat production rate from radioactive decay for the upper crust as 3.73 μWm^{-3} by U, Th and K concentrations in the volcanic and plutonic rocks of the western Anatolia given by Ercan et al. (1983). Assuming the characteristic depth (h_r) to be 13-15 km, IlkiSik (1995) also suggested the crust's contribution to the surface heat flow and the reduced heat flow as 52 and 55 mWm^{-2} respectively, based on the linear relationship between the near surface heat production and surface heat flow.

As to the magnitude of the mean heat flow value in the western Anatolia, both Tezcan and Turgay (1991) and IlkiSik (1995) calculated very high main values, about 120 and 107 mWm^{-2}, respectively. They both used the measurements taken only in geothermal systems, and this has likely affected the magnitudes because hydrothermal circulation plays an important role in the geothermal systems. This circulation might produce high near surface heat flow, which may not represent the deeper heat flow and main values of the region.

These studies reveal that the western Anatolia has anomalous heat flow compared to the rest of the country. Since the heat flow data are insufficient, to explain the cause of such high heat flow is problematic. A generally accepted hypothesis is that there is a crustal thinning in western Anatolia as a result of the extensional tectonic regime, and this also formed a quite shallow asthenosphere under western Anatolia. Therefore, the asthenosphere should be the main source of anomalous heat flow in the region (Alptekin et al., 1990; IlkiSik, 1995). On the other hand, Tezcan (1995) suggests an association between high heat flow anomalies and metamorphic massifs (the Menderes, Kazdag, KirSehir massifs, etc.). He considers high content of radioactive elements in massifs as the source of the high heat flow.

CRUSTAL STRUCTURE

Modeling temperature distribution within the crust requires the knowledge of crustal structure, thermal properties of the geological units and some boundary conditions. Even though the crustal structure in western Anatolia is of great interest, there are no data from explosion seismology that can be employed to construct a reliable crustal structure (Alptekin et al., 1990). Recently, Saunders et al. (1998) investigated the crustal structure of western Turkey using teleseismic receiver functions. They used the data at two locations, Kula and USak. They estimated the crustal thicknesses at these locations as ~30 and ~34 km, respectively. While the crustal thickness increases from Kula to USak, the effects of the extension in the region decrease. Makris (1978) used topography, gravity and limited seismic refraction data in the Aegean Sea and estimated a crustal thickness ranging from about 22 km in the central Aegean Sea to over 40 km in the Anatolian Plateau. Mindevalli and Mitchell (1989) give an average crustal thickness of about 40 km which is thinner on the west coast (~34 km) using fundamental-mode Rayleigh and Love wave group velocities for western Turkey. Also Ezen (1991), based on Rayleigh wave dispersion, reported similar result (~30-32 km thick) for the crust in the western Anatolia.

Gravity data can indicate the broad features of the crust and upper mantle structure. Therefore, the Bouguer gravity map of Turkey published by MTA (1979) on the scale of 1/500000 were used to model crustal structures along the Profiles I and II (Fig. 1). The profiles were sampled using a 5-km sampling interval. Appropriate forward modeling was used for constructing crustal models (Talwani et al., 1959). The method is based on an evaluation of gravity anomalies caused by bodies of arbitrary shapes. The method performs a double integration analytically and a single integration numerically. The body whose anomaly has to be evaluated is first represented by contours. Each contour is then replaced by a polygonal lamina. By making the number of sides of this polygon larger, the contour boundary can be represented as accurately as desired. The numerical integration is then carried out with respect to Z axis (for details, see Talwani, 1965).

Generally, rocks making up the upper parts of the crust tend to be felsic in composition and have densities between 2.6 and 2.8 g/cm^3. Rocks in the lower part of the crust are more mafic in composition and their densities range from 2.8-3.0 g/cm^3. Mantle is mainly composed of ultramafic rocks with the density of 3.2 g/cm^3 approximately (Dobrin and Savit, 1988).

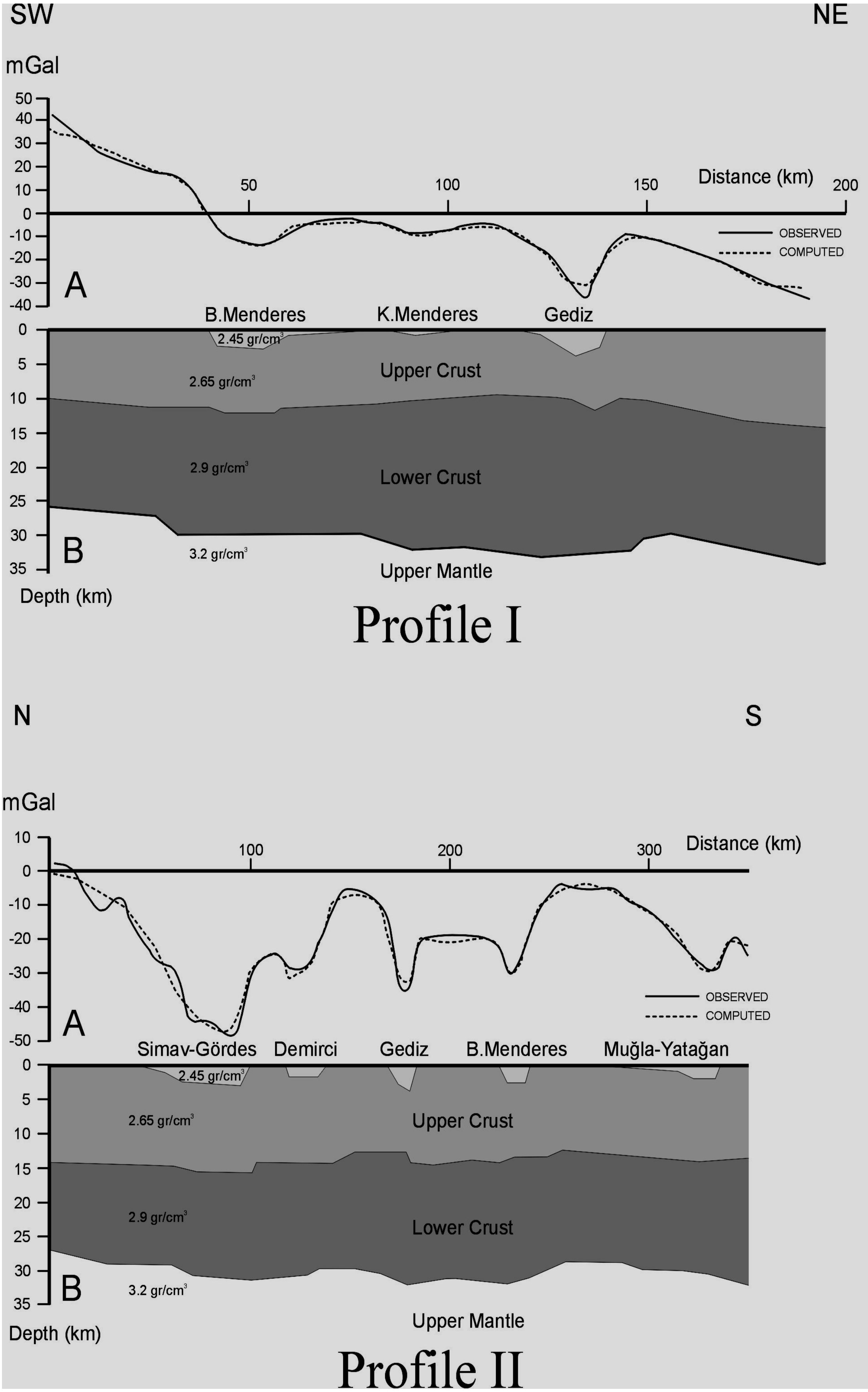

FIG. 3.. Crustal structure used in the thermal modeling. Gravity data (A) and the model (B) along the Profile I (top); gravity data (A) and the model (B) along the Profile II (bottom).

Therefore, considering the geology and rock distribution in the region, we assigned the density values of 2.45, 2.65, 2.9, 3.2 g/cm^3 for the sedimentary fill in the grabens, upper and lower crust and upper mantle, respectively during the forward modeling of the gravity data.

Figure 3 shows the gravity data and crustal models along the profiles. The length of the Profile I is approximately 200 km. It crosses the Büyük Menderes, Küçük Menderes and Gediz grabens. Low gravity values exist at the grabens. The length of the Profile II is about 350 km. This profile includes the Simav-Gördes, Demirci subgrabens, the Gediz, Büyük Menderes grabens and Mugla-Yatagan subgrabens. Again, the grabens are characterized by low gravity values. On the other hand, there is a gradual increase in the Bouguer gravity anomalies toward the west along the Profile I indicating a relative thinning of the crust in this direction whereas no such thing could be observed along the Profile II running in the north-south direction. This fits very well with the regional setting of the Aegean domain where there is a back-arc uplifting behind the Hellenic arc.

According to geological data, grabens in the western Anatolia have asymmetric forms (Yilmaz et al., 2000; Sözbilir, 2001). Our forward modeling results also indicate such a feature. In our models, the depths of the grabens range from 2.5 to 3.5 km; they are asymmetric; the average thickness of the crust is about 30 km (Makris, 1978; Mindevalli and Mitchel, 1989; Ezen, 1991; Saunders et al., 1998). Paton (1992) also studied Bouguer gravity data over the Büyük Menderes and Gediz grabens and estimated an average depth to the basement as ~1.5-2 km and Gürer et al. (2002) suggest a varying sediment thickness in the range of 1-3.8 km along the Gediz graben based on geoelectrical data (mainly magnetotelluric data). Since there are no deep seismic studies in the study area, the knowledge of the sediment thickness in the grabens is not satisfactory. Even though there are wells drilled for geothermal explorations they do not give enough information for sediment thickness because they are located close to the borders of the grabens.

METHOD OF THERMAL MODELING

To calculate temperature field in the crust a two-dimensional steady-state heat conduction equation was numerically solved using successive overrelaxation method of finite-differences. The two-dimensional steady-state heat conduction equation is given as the following:

$$\frac{\partial}{\partial x}\left(k\frac{\partial T}{\partial x}\right)+\frac{\partial}{\partial z}\left(k\frac{\partial T}{\partial z}\right)+A=0 \qquad (1)$$

where k(x,z) is thermal conductivity; T(x,z) is temperature field; A(x,z) is heat production; x and z are coordinates (Cermák et al., 1991).

Following Kukkonen and Jõeleht (1996), we took into account the temperature dependence of thermal conductivity (Wm^{-1}K^{-1}), including the effects of both lattice conductivity and radiative heat transfer as:

$$k=\frac{k_o}{(1+aT)}+b(T+273.15K)^3 \qquad (2)$$

Here, k_o is the thermal conductivity (Wm $^{-1}$ K^{-1}) in the surface conditions (20°C), T is the temperature (°C), a (K^{-1}) and b (Wm^{-1}K^{-4}) are the experimental parameters. For *a* the value of 0.0015 was assigned approximating the lattice conductivity of common rock types (Zoth and Haenel, 1988) for temperatures lower than 1000-1200°C. The radiative heat transfer becomes important in the lower crust and mantle at temperatures beyond 800°C. The *b* value was assumed as $1\cdot10^{-10}$ (Schatz and Simmons, 1972) for ultrabasic mantle material. The variation of thermal conductivity with pressure has been ignored since its effect on the conductivity is small compare to temperature's effect. Also, how thermal conductivity changes with pressure is not well documented (Sams and Thomas-Betts, 1988).

For heat production, we assumed an exponential decrease with depth for radioactivity in the crust because it is one of the available models (step and linear models) satisfying the linear heat flow relation (Lachenbruck, 1970). Also it is the model which represents well the actual situation in the continents (Condie, 1976). This model is defined as follows:

$$A(z) = A_0 \exp\left(\frac{-z}{h_r}\right) \qquad (3)$$

where A(z) (μWm^{-3}) is the heat production at depth z; A_o is the near-surface heat production (μWm^{-3}); h_r is the characteristic depth (km), and z is the depth (km). For A_o we assigned the value of 3.73 μWm^{-3} (IlkiSik, 1995). h_r is taken as 12 km for the western Anatolia considering the average thickness of the upper crustal layer (granitic-gneissic part of the crust) in the models. The depth z ranges from 0-35 km. As to the heat generated by radioactive decay, no heat production in the grabens was considered, i.e. the heat production equals to zero everywhere in the grabens.

Solution of equation (1) is subject to the following boundary conditions:

- T(x, z=0) = 18ºC, which is considered as the mean annual surface temperature (Tezcan, 1992),
- No-flux boundary conditions are applied to the vertical boundaries of the models, i.e. $\left(\frac{\partial T}{\partial x}\right) = 0$ at x=0 and x=L, L is the length of the model,
- A constant vertical heat flow is assigned along the lower boundaries of the models. Considering the linear relationship between the surface heat flow and near-surface heat production, the reduced heat flow is the heat coming from the mantle and lower crust (Condie, 1976). Therefore, integrating the heat production function A(z) from h_r to 35 km yields the lower crust's contribution. Subtracting this from the reduced heat flow should give the contribution of mantle, which can be used as lower boundary condition; and this is about 30 mWm^{-2}.

Solving equation (1) requires a certain crustal model and thermal conductivity values for the geological units (Cermák et al., 1991). Table (1) shows the units and corresponding thermal conductivities in surface conditions (k_o).

Table 1

Units (Wm^{-1} K^{-1}) Reference	k_o
Sedimentary fill in grabens (Tezcan and Turgay, 1991)	2.1
Upper crust (Cermák et al., 1991)	3.0
Lower crust (Cermák et al., 1991)	2.0
Upper Mantle assumed	2.8

RESULTS AND DISCUSSION

Since we assign a constant heat flow value to the lower boundary of the models and the topographies of the model interfaces change smoothly, relatively flat isotherms are obtained. Thus, we preferred to plot temperatures at different depths versus distance to reveal spatial temperature changes. Figure 4 and Figure 5 show the calculated temperatures along the Profiles I and II, respectively. These figures indicate that temperature distributions in the upper and lower crust are remarkably different. The distributions in the upper crust exhibit undulating variations, whereas the ones in the lower crust show some smooth spatial changes.

Figure 4 shows the temperatures at the depths of 1, 5, 10, 15 and 30 km along the Profile I. Considering the depth range of the grabens in the models, Figure 4a displays the temperature change along a depth level cross-cutting the Büyük Menderes, Küçük Menderes and Gediz grabens. According to this plot, temperatures outside the grabens are approximately 55°C while they are about 65°C inside the grabens. In other words, temperatures at 1 km in the grabens are about 10°C higher than those in the same depth in the surrounding region. At the depth of 5 km (Fig. 4b), we see a similar situation even though the grabens do not extent down to this depth. The regions under the Büyük Menderes and Gediz grabens are about 20°C, and the region under the Küçük Menderes graben is about 10°C hotter than the rest of region. This can be explained considering the thermal conductivities of the various units. Because the grabens are filled by clayey sediments (Tezcan, 1995) whose thermal conductivities are as low as 2.1 $Wm^{-1}K^{-1}$, they cannot conduct the heat from below as fast as the surrounding region which has relatively high thermal conductivity (3.0 $Wm^{-1}K^{-1}$). Because of this, the grabens behave like thermal insulators, which do not let heat flow transfer easily through the surface, and this causes the grabens and the regions under them to become relatively hotter compared to the surrounding regions.

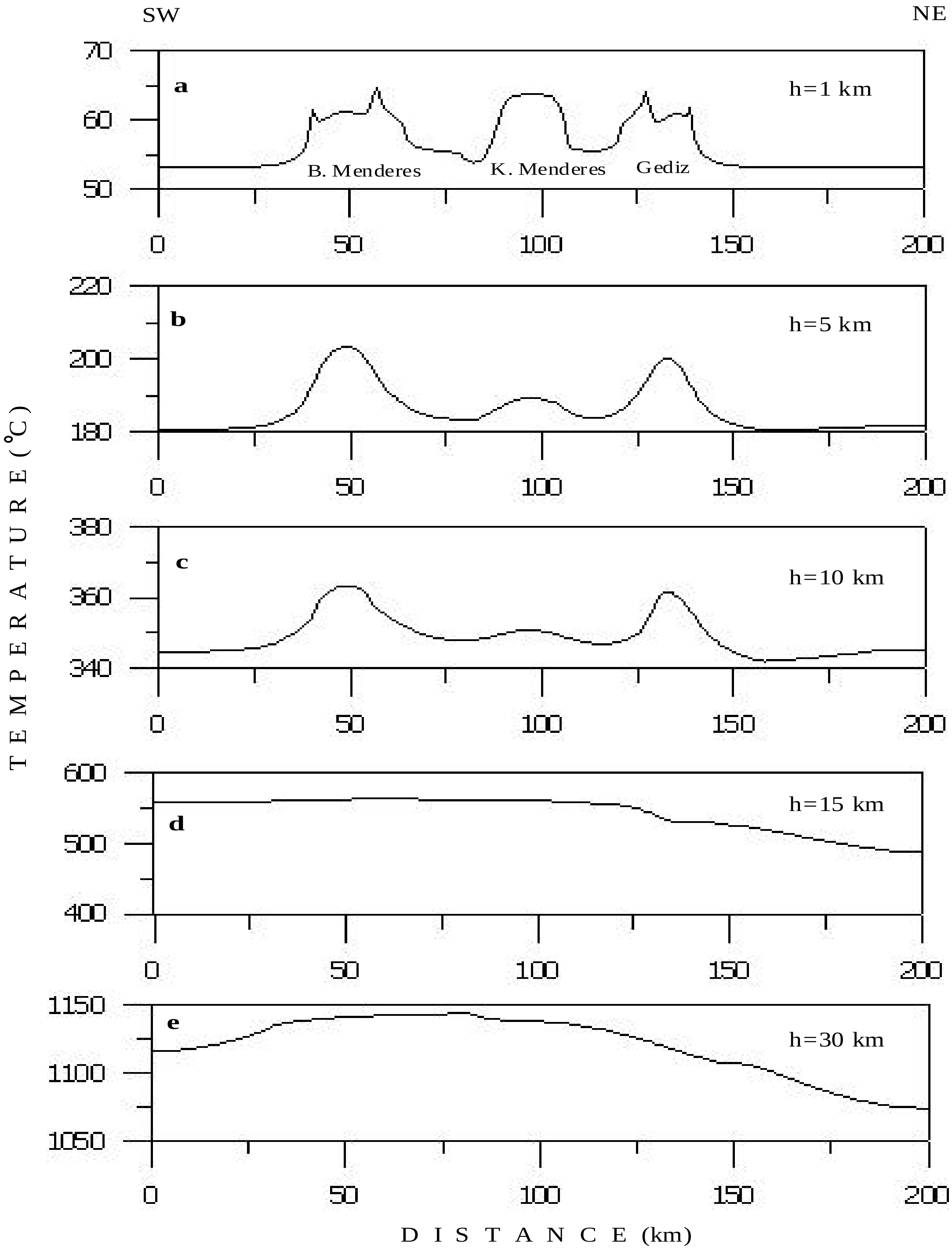

FIG. 4. Modelled temperatures at various depths along the Profile I. Plots show temperature change with distance at the depths of 1, 5, 10, 15, 30 km.

Temperatures at 10 km (Fig. 4c) range from 345-360°C and they present variations which are similar to Figures 4a and 4b. But the effect of the Küçük Menderes graben on the temperatures is not as strong as the ones of the Büyük Menderes and Gediz grabens. It seems that the region under the Küçük Menderes graben is a few degrees warmer than the surrounding region. The depth extent of the graben can explain this, as the depth of the Küçük Menderes graben is quite shallow and it has less sedimentary fill. Thus, it cannot affect the deeper regions as much as the Büyük Menderes and Gediz grabens.

Usually, parts of the crust down to the depths where 300±100°C isotherms exist are considered to be brittle enough to nucleate an earthquake (Chen and Molnar, 1983). Generally, brittle-ductile transition in the crust is assumed to be at the depth where 350°C isotherm takes place (Kukkonen and Jõeleth, 1996); and the brittle part of the crust is considered as seismogenic zone. As to the temperature range in Figure 4c (from 345 to 360°C), we can conclude that the seismogenic zone is about 10 km thick along the Profile I. This result is also in accordance with seismological observations. Jackson and McKenzie (1988) suggest a 10-15 km thick seismogenic zone in the western Anatolia; and Eyidoðan and Jackson (1985) has similar result considering the focal depths of the largest normal faulting earthquakes which are about 6-10 km in the region, and this implies that the seismogenic zone should be about 10 km thick.

A temperature at which a magnetic mineral loses its ferromagnetism is called as the Curie temperature. Even though the magnitude of the Curie temperature depends upon the mineralogy of the rocks, it ranges in a narrow zone, which is about 520-560°C within the most part of the continental crust. Also, the Curie isotherm defines the base of the magnetic crust (Mayhew, 1982). Figure 4d shows temperatures of this order at 15 km. They range from 500 to 560°C. They are almost fixed at the value of 560°C up to the distance of 130 km from the beggining of the profile, and they start to decrease gradually to the north of this point, reaching 500°C at the end of the profile. Based on this, it is concluded that the thickness of the magnetic crust along the Profile I should be around 15 km.

In our crustal models, the average thickness of the crust in western Anatolia is about 30 km. Figure 4e depicts the temperatures at the depth of 30 km, they can be regarded as the temperatures in the vicinity of Moho discontinuity, i.e. temperatures at the bottom of the crust. According to this figure, they range approximately from 1075 to 1145°C. Temperatures at the southwest end of the model are about 1100°C, increasing towards the northeast and reaching around 1145°C at the distance of 80 km. After this point they start to gradually decrease, and at the northeast end of the model temperature is about 1075ºC. We can therefore conclude that the temperature at the bottom of the crust along the Profile I is about 1100ºC.

Figure 5 shows temperatures at depths of 1, 5, 10, 15 and 30 km along the Profile II. Like the temperatures along the Profile I, there is noticeable difference between the temperatures in the upper and lower crust. Similar to the Profile I, the distribution in the upper crust shows undulating variations, whereas the ones in the lower crust smoothly change with distance.

Figure 5a displays the temperatures at the depth of 1 km, that is, a depth level cross-cutting the Simav-Gördes, Demirci, Gediz, Büyük Menderes and Mugla-Yatagan grabens. Like the situation along the Profile I, the grabens have relatively higher temperatures, about 65ºC inside the grabens and 55ºC in the surrounding region, i.e. the temperatures at 1 km in the grabens are about 10ºC higher those at the same depth in the surrounding region. Because of the geometry of the Mugla-Yatagan graben temperature gradually rises up to 65ºC toward the south. Temperatures at 5 km (Fig. 5b) have similar character but increased magnitudes. The highest temperature (about 210°C) is obtained in the Simav-Gördes graben. The other grabens reach temperatures of 200ºC. This shows that the regions under the grabens is approximately 20-25ºC hotter than the surrounding region, which is about 180ºC.

The variation of temperature at 10 km (Fig. 5c) is very similar to that of Figure 5b, with temperatures ranging from 340-370ºC. Based on these results we can conclude that the grabens and the regions underneath them have higher temperatures than the surrounding regions along the Profile II. As mentioned above, this is explained by the effect of the sedimentary fill in the grabens and heat transfer by hot waters uprising from deeper sites along extension fault system.

According to temperatures in the Figure 5c, the seismogenic zone along the Profile II is also about 10 km thick. Figure 5d shows the results at 15 km, where temperatures range from 500 to 525ºC and do not show significant changes. These values imply that the depth to the Curie isotherm, like the Profile I, should be also about 15 km along the Profile II. Figure 5e

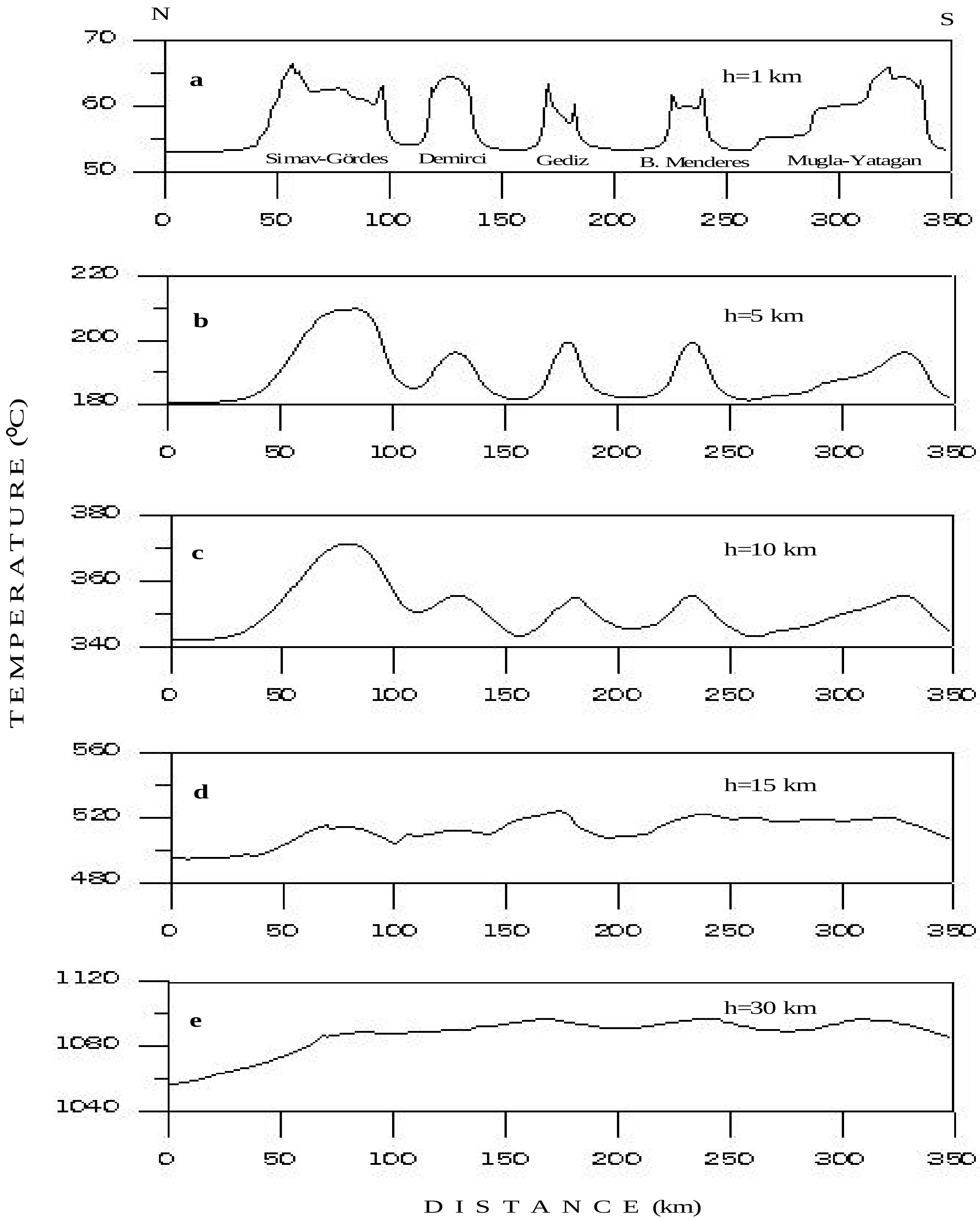

Figure 5. Modelled temperatures at various depths along the Profile II. Plots show temperature change with distance at the depths of 1, 5, 10, 15, 30 km.

corresponds to temperatures at 30 km which are considered temperatures in the vicinity of Moho discontinuity. They start around 1060ºC at the north end of the profile, rising to 1080ºC at the distance of 70 km. After this point, temperature undulates between 1100 and 1080ºC until the distance of 320 km and then decreases gradually down to 1080ºC at the south end of the profile. Therefore, it can be concluded that the temperatures at the bottom of the crust along the Profile II is about 1075ºC.

CONCLUSION

Based on the numerical modeling of two-dimensional steady-state conductive heat transfer in western Anatolia, the results obtained show that the temperatures in the grabens and regions under them are relatively higher than those in the surrounding regions; the thickness of the seismogenic zone and depth to the Curie isotherm were estimated as 10 and 15 km, respectively. Due to the sedimentary fill in the grabens they affect the temperature distributions in the upper crust. Because of their low thermal conductivities they behave like a thermal insulator. This causes the grabens and regions which extend down to about 10 km under them to become 10-20°C hotter than the surrounding region. Finally, the temperature at the bottom of the crust was calculated as 1075-1100ºC approximately.

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