Geoscience Frontiers 4 (2013) 389–398

Contents lists available at SciVerse ScienceDirect

China University of Geosciences (Beijing)

Geoscience Frontiers

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

Research paper

Geothermal structure of the eastern Black Sea basin and the eastern Pontides orogenic belt: Implications for subduction polarity of Tethys oceanic lithosphere

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A R T I C L E I N F O

Article history:
Received 9 July 2012
Received in revised form
19 January 2013
Accepted 1 February 2013
Available online 16 February 2013

Keywords:
Eastern Black Sea basin
Eastern Pontides
Heat flow
Lithospheric thickness
Geodynamic evolution
Subduction polarity

A B S T R A C T

The numerical results of thermal modeling studies indicate that the lithosphere is cold and strong beneath the Black Sea basin. The thermal lithospheric thickness increases southward from the eastern Pontides orogenic belt (49.4 km) to Black Sea basin (152.2 km). The Moho temperature increases from 367 °C in the trench to 978 °C in the arc region. The heat flow values for the Moho surface change between 16.4 mW m^{-2} in the Black Sea basin and 56.9 mW m^{-2} in the eastern Pontides orogenic belt. Along the southern Black Sea coast, the trench region has a relatively low geothermal potential with respect to the arc and back-arc region. The numerical studies support the existence of southward subduction beneath the Pontides during the late Mesozoic–Cenozoic.

1. Introduction

Geophysical and geochemical data show that the arc and back-arc regions are hot due to the cooling effect of the subducting plate (Watanabe et al., 1977). The highest heat flow values associated with melting processes, are observed over the volcanic arc. The lowest heat flow over the subduction complex and fore-arc basin is assumed to be the result from thermal blanketing effect by the overlying plate and accreted sediments above the subducting oceanic plate (Langseth et al., 1990). The zone from arc to trench axis shows less than normal heat flow values due to increased hydrothermal circulation resulting from flexing and normal faulting of the plate prior to subduction (Yamano and Uyeda, 1990; Langseth and Silver, 1996). Numerical modeling in elucidating the characteristics of convergent margin tectonics has been attempted in a number of recent studies (e.g., Baitsch-Ghirardello et al., 2012; Sizova et al., 2013; Vogt and Gerya, 2013).

The Black Sea is a marginal sea located in the Alpine folded belt, represented by the Caucasus to the east, by the Balkanides–Pontides to the south and southwest, the Crimean Mountains to the north, and young platforms (Moesian and Scythian) to the west and northwest (Fig. 1). The geological setting of the basin has been known for many years (Ross et al., 1974; Letouzey et al., 1977; Zonenshain and LePichon, 1986; Manetti et al., 1988; Okay et al., 1994). The Black Sea basin is divided into two parts, western and eastern, separated by the Mid-Black Sea ridge (Shevchenko and Rezanov, 1972; Datchev, 1977; Tugolessov et al., 1985; Chekunov, 1987; Finetti et al., 1988).

The most acceptable hypothesis about the mechanism which caused the origin of the Black Sea, suggests that the Black Sea is a back-arc basin which was opened behind the Pontide magmatic arc and formed in the process of continental crust taphrogeny and spreading (Adamia et al., 1974; Okay et al., 1994). The duration of formation of this basin is estimated as from Paleozoic to Mesozoic and Cenozoic, including Neogene and Anthropogene (Zonenshain and LePichon, 1986; Belousov and Volvovsky, 1989). The mechanisms of the origin and evolution of the Black Sea basins are debated. Also, the age of the western and eastern basins remains controversial.

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Peer-review under responsibility of China University of Geosciences (Beijing)

Production and hosting by Elsevier

1674-9871/$ – see front matter
http://dx.doi.org/10.1016/j.gsf.2013.02.001

Figure 1. Tectonic elements of the Black Sea basin and the Anatolian plates (Robinson et al., 1996). The rectangle shows the study area.

The eastern Pontides orogenic belt consists of a mountain chain extending along the southeast Black Sea coast with a length of 500 km and width of 200 km and is subdivided into Northern, Southern and Axial zones from north to south based on different lithological units, facies changes and tectonic characteristics (Bektaş et al., 1995; Eyuboglu et al., 2006). The Northern zone is characterized by Mesozoic–Cenozoic volcanic rocks and granitic intrusions. The Cenozoic rocks are typically represented by Eocene mafic volcanic rocks and granitic intrusions located in the Northern zone. The Southern zone includes numerous rock associations such as Paleozoic metamorphic belts and granitic intrusions, Pre-Liassic Alaskan-type mafic–ultramafic intrusions, Mesozoic–Cenozoic sedimentary sequences, late Cretaceous shoshonitic and ultra-potassic magmatic units and late Paleocene–early Eocene adakitic intrusions. Upper mantle peridotites and middle to upper Cretaceous olistostromal melange are extensively exposed farther south in the Axial zone (Eyuboglu et al., 2007; Eyuboglu, 2010).

The geodynamic evolution of the eastern Pontides orogenic belt has been debated, primarily due to the scarcity of structural, geochemical and geochronological data. Various models proposed for the geodynamic evolution of the eastern Pontides orogenic belt can be grouped into three. Some workers propose that the ultramafic rocks exposed to the south of the Pontide arc represent the remnants of the Paleotethys Ocean (Adamia et al., 1977; Ustaömer and Robertson, 1997), and suggest that the eastern Pontides orogenic belt developed by northward subduction of Paleotethys, which was situated south of the magmatic arc, from the Paleozoic until the end of Eocene. Recent studies performed by Topuz et al. (2005), Dilek et al. (2010) and Karslı et al. (2010) proposed that the early Cenozoic magmatism in the eastern Pontides magmatic arc was generated by partial melting or delamination of thickened lower continental crust after the collision between the Pontides and the Tauride platform during the northward subduction of the Neotethys oceanic lithosphere in the Paleocene–early Eocene.

In contrast, Şengör and Yılmaz (1981) proposed that the Paleotethys was situated at the north of the Pontides, hence requiring southward subduction from the Paleozoic until the mid-Jurassic, followed by northward subduction from the upper Cretaceous until the end of the Eocene. According to their model, the eastern Pontides orogenic belt represents the southern active continental margin of Eurasia, along which the late Cretaceous arc volcanism developed above the northward subducting Neotethys. In this model, the Neotethys was interpreted as a back-arc basin to the eastern Pontides magmatic arc that had originally opened in the lower Jurassic.

The third model includes those of Dewey et al. (1973), Chorowicz et al. (1998), Bektaş et al. (1999), Eyuboglu et al. (2006, 2007) and Eyuboglu (2010) who argued that southward subduction continued uninterruptedly from the Paleozoic till the end of Eocene. According to this model, the Black Sea is a remnant of the Paleotethys Ocean and the ophiolitic belt represents a back-arc basin environment. Also, Eyuboglu et al. (2011a,b,c,d, 2012a, 2013) proposed that the eastern Pontides were formed by slab window processes in a southward subduction zone based on systematically geological, geochemical and isotopic data obtained from the late Paleocene–early Eocene adakitic rocks.

The aim of this study is to interpret the surface heat flow density and crustal structure data and to construct the thermal structure of the lithosphere beneath the eastern Pontides orogenic belt and the eastern Black Sea basin to explain the geodynamic evolution of the study region in terms of the subduction polarity.

2. Previous studies on the crustal and lithospheric structure of the Black Sea basin and eastern Pontides orogenic belt

The main features of the structure inferred from the refraction seismic profiling data performed by Soviet geophysicists in 60–80's are a lack of the granitic layer in the central part of the Black Sea (Mindeli et al., 1965; Malovitsky and Neprochnov, 1966; Boulange et al., 1975; Bulanzhe et al., 1975; Tugolessov et al., 1985; Sollogub, 1986; Yanovskaya et al., 1998). The depth of Moho is about 20 km in the center of the western Black Sea basin. The Moho reaches to a depth of 45 km beneath the Pontides mountain belt and to a depth of 40–45 km beneath the Russian Platform. On the other hand, the Moho depth rises to about 25 km under the eastern Black Sea basin. The sediment thickness (from Mesozoic to the present) in the West

Black Sea basin reaches 16–18 km and in the East Black Sea basin 12–14 km. According to the study of Belousov and Volvovsky (1992), it is determined that the crustal thickness is 18–19 km in the western and 23 km in the eastern basins. Bulanzhe et al. (1975) showed that the deep part of the Black Sea basin has a crustal velocity of 6.8 km s^{-1} indicating the oceanic crust by the DSS data.

The lithosphere thickness in the western and eastern Black Sea basins was determined as 80–90 km (Golmshtok and Hahalev, 1987; Golmshtok et al., 1992). Spadini et al. (1996, 1997) performed thermomechanical modeling of the Black Sea basin and suggested that the western Black Sea basin was formed by rifting of thick and cold lithosphere (200 km), the eastern Black Sea basin was developed by rifting of thin and warm lithosphere (80 km). Meredith and Egan (2002) represented the temperature at the base of lithosphere (125 km) as 1333 °C throughout the basin's post-rift stage corresponding to the development of a more thermally mature lithosphere. According to the Verzhbitsky (2002) the lithospheric thickness of the western and eastern Black Sea basins calculated from heat flow data (60–65 km) corresponds to the thickness of the early Cenozoic oceanic lithosphere.

Starostenko et al. (2004) estimated that the depth of Moho is 19 km beneath the western Black Sea basin and is 22 km beneath the eastern Black Sea basin with the gravity studies. The deepest Moho (40 km) is found beneath the Shatsky Ridge. The Moho depths are determined as 28 and 34 km in the Tuapse and Sorokin troughs, respectively. Beneath the Mid-Black Sea Ridge, the base of the crust forms troughs of 29 and 33 km in the south and north, respectively. Also, they found a two density differences (-36 kg m^{-3} and $+27$ kg m^{-3}) at the depth interval of 25–160 km and 200–250 km for the Precambrian cratonic lithosphere reference.

Several workers have studied the crustal structure of the eastern Anatolia and eastern Pontides including Mindevalli and Mitchell (1989), Çakır et al. (2000), Seber et al. (2001), Al-Lazki et al. (2003), Gök et al. (2003, 2007, 2011), Şengör et al. (2003), Zor et al. (2003), Çakır and Erduran (2004), Sandvol and Zor (2004), Angus et al. (2006), Barazangi et al. (2006), Pamukçu et al. (2007) and Maden et al. (2009a,b). Mindevalli and Mitchell (1989), who indicated possibility of anisotropy in the upper crust, modeled the crust and upper mantle velocity structure using Rayleigh and Love-wave group velocities and computed a 40 km crustal thickness beneath the eastern Turkey. Çakır et al. (2000) modeled the crustal structure by using teleseismic three-component digital data from Trabzon. They found that the Moho dips approximately southwards and thickens from 32 to 40 km beneath the eastern Pontides orogenic belt.

According to Seber et al. (2001), the Turkish–Iranian plateau is underlain by a crust of 45–50 km, which is thicker than average continental crust, consistent with the evidence for post late middle Miocene crustal shortening. Low Pn velocities and high Sn attenuation (Al-Lazki et al., 2003; Gök et al., 2003) indicate that lithospheric mantle is either thinned or totally removed beneath the eastern Turkey. The crustal thicknesses are found to be thinner (45–48 km) relative to its high elevation of the plateau. In the East Anatolian High Plateau, Şengör et al. (2003) revealed a hot mantle but not a thick crust of 45 km by using seismic data gathered from 29 seismograph stations network. Zor et al. (2003) examined the crustal structure of the eastern Anatolian plateau and estimated an average depth of approximately 45 km, thickening from west to east and south to north and an average crustal shear-wave velocity of 3.7 km s^{-1}.

Çakır and Erduran (2004) inferred that Moho discontinuity is located at ca. 35 km depth by using teleseismic radial-component receiver functions and regional Rayleigh and Love-wave group velocities beneath TBZ station located on the northern side of the eastern Pontides orogenic belt. Angus et al. (2006) found that the lithosphere–asthenosphere boundary beneath eastern Anatolian accretionary complex remains shallow and is approximately parallel to the Moho with an average depth of roughly between 30 and 55 km, which suggests the velocity contrast between the lithosphere and asthenosphere is small possibly due to a small temperature increment, under the 39° and 40° N sections.

Another research carried out by Barazangi et al. (2006) depicts the average thickness of the crust as ca. 45 km utilizing the gravity data in eastern Turkey. Gök et al. (2007) evaluated average crustal thicknesses of 44 km in the Anatolian Block and 48 km in the Anatolian plateau. They also observed that the lithospheric mantle has been thinned or completely removed underneath the Anatolian plateau. On the other hand, Pamukçu et al. (2007) found the crustal thickness changes from 38 to 52 km in the eastern Anatolia.

Maden et al. (2009a) investigated the two-and-three dimensional crustal structure of the eastern Pontides orogenic belt from gravity data. The average Moho, Conrad and basement depth of the eastern Pontides orogenic belt are determined by using spectral analysis technique as 35.7, 26.5 and 4.6 km, respectively. Also, they presented that the Moho depths vary from 33.9 to 42.6 km by using power spectrum method and change between 30.1 and 43.8 km determined from gravity inversion studies. Maden et al. (2009b) determined the crustal and tectonic structure of the eastern Pontides orogenic belt by using gravity and magnetic data. They constructed a Moho depth variation map by using the power spectrum method applied to the gravity data and concluded that the Moho depth for the eastern Pontides orogenic belt varies from 29.0 km in the north to 47.2 km in the south. The map of the Moho depths beneath the eastern Pontides and the eastern Black Sea basin is seen in Fig. 2a. Another study by 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.

3. Previous studies on the thermal structure of the Black Sea basin and eastern Pontides orogenic belt

Heat flow values within the Black Sea basin are complicated and difficult to interpret because of the thick sedimentary infill. The characteristic features are heat flow density anomalies running transverse to the main course of the Alpine units. They are related to deep fault zones crossing these structures (from the Arabian plate to peri-Caucasus region). A genetic relationship can be drawn from the analysis of the heat flow map of the study region between the heat flow pattern and tectonic structure of the Black Sea and eastern Pontides orogenic belt which is located between the converging Eurasia and Arabian–African lithospheric plates within the wide zone of continent–continent collision (Kutas et al., 1998).

Whereas low heat flow density values dominate in the Black Sea basin, the eastern Pontides orogenic belt shows relatively higher heat flow values. The heat flow values rapidly increase from oceanic to the continental crust (Fig. 2b). The highest heat flow values (105 mW m^{-2}) associated with extensive Neogene and Quaternary volcanic activity are observed in the eastern Pontides orogenic belt. The lowest heat flow values (12 mW m^{-2}) are seen in central part of the eastern Black Sea basin where the maximum sediment thickness is seen. According to Nikishin et al. (2003), surface heat flow values vary between 30 mW m^{-2} in the center of the basin and 70 mW m^{-2} in the Crimea and Caucasus margins of the basin. The surface heat flow pattern in study region shows that heat flow in the eastern Pontides orogenic belt (stable block) is distinct from that in eastern Black Sea basin.

Maden (2012a) computed Moho temperature of 590 ± 60 °C at a depth of 35 km indicating the presence of a brittle–ductile transition zone in the eastern Pontides orogenic belt. This temperature value might be related to water in the subducted crust during southward-dipping subduction of the Tethys oceanic lithosphere. Moreover, the depth of Curie temperature which is above the Moho

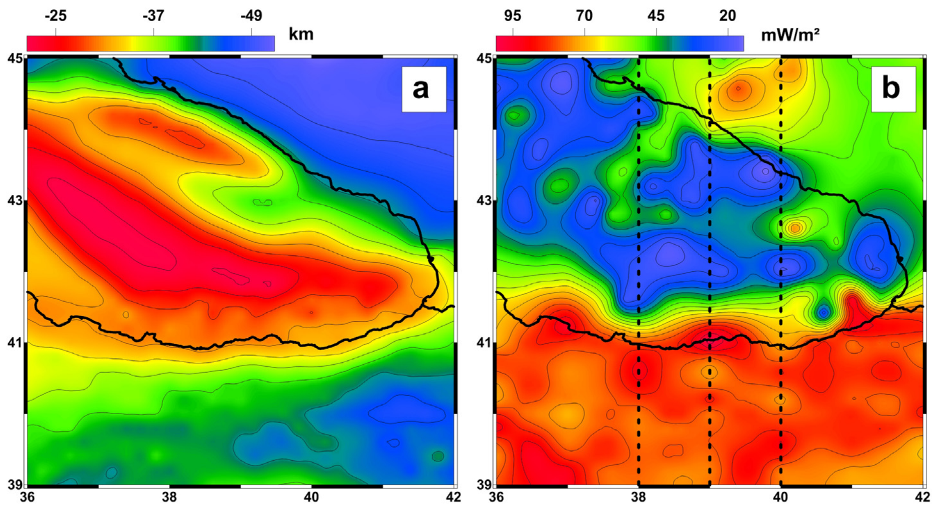

Figure 2. (a) Crustal structure map of the eastern Black Sea basin and eastern Pontides orogenic belt (Starostenko et al., 2004; Maden et al., 2009a). (b) Surface heat flow density map of the study area (modified from Kutas et al. (1998), Verzhbitsky (2002) and Maden (2012a), the cutted lines demonstrate the profile locations).

depth with the maximum difference of 5–7 km is found at 29 km, which is more or less consistent with the Curie temperature depth value given by Aydın et al. (2005) and Maden et al. (2009b). While the surface heat flow density values are estimated between 66.5 and 104.7 mW m^{-2}, the mantle heat flow density value, which is related to melting of the lithospheric mantle caused by upwelling of the asthenosphere, is obtained for the area as 48 mW m^{-2}.

Maden (2012b) performed 2D thermal modeling studies along the central Pontides magmatic arc (northern Turkey), Sakarya and Kırşehir continents in order to delineate the crustal thermal structure and subduction polarity. According to this study, Moho temperatures in the region were found between 992 °C in the south (back-arc) and 415 °C in the north (arc). Moreover, mantle heat flow values vary from 57.2 mW m^{-2} in the south (back-arc) to 34.7 mW m^{-2} in the north (arc). Maden (2012b) delineated that the Eurasia plate had moved from north to south under the Anatolia plate along the south Black Sea coast.

Three surface heat flow profiles, the locations of which are shown in Fig. 2b, beneath the 38°, 39° and 40° latitudes along the NS direction, are illustrated in Fig. 3a–c. These profiles depict a typical surface heat flow pattern over subduction zones through trench to the back-arc region. In these figures, the surface heat flow values are low in the trench and high in the arc and back-arc regions.

4. Geothermal modeling

Temperature is the most important factor for determining the rheological structure of the lithosphere. Thermal structure of the lithosphere can be approximated by the steady state solution of the heat conduction equation:

$$\nabla^2 T = -\frac{A}{k} \tag{1}$$

where T is the temperature (°C), A is the radioactive heat production in μW m^{-3}, k is the thermal conductivity in W m^{-1} K^{-1}. The solution of this partial differential equation permits the calculations of stable geotherms within the crust and lithospheric mantle. Such calculations are constrained primarily by surface heat flow measurements, thermal conductivity and heat production within the crust and the lithospheric mantle. To calculate heat flow and temperature in the lithosphere a well-founded distribution of radiogenic heat sources and thermal conductivity within the depth range is needed (Artemieva and Mooney, 2001).

Throughout the crust, thermal conductivity varies with the composition and with both the pressure and the temperature. The thermal conductivity of most crustal rocks varies inversely with the temperature and tends to increase with the depth or the increasing pressure according to the following equation:

$$k = \frac{k_0(1+bz)}{1+cT} \tag{2}$$

where k_0 represents the thermal conductivity at surface conditions, b and c are experimental constants which control the behavior of k depending on the lithology (Chapman and Furlong, 1992). The values of k_0, c and b are constant in each layer of the crust. T is temperature in °C. In the upper crust, $k_0 = 3.0$, $c = 0.0015$, $b = 0.0015$; in the lower crust, $k_0 = 2.6$, $c = 0.0015$, $b = 0.0001$ (Chapman, 1986).

Correia and Ramalho (1999) used the thermal conductivity values for the upper, the middle and the lower crust as 2.7, 2.5 and 2.1 W m^{-1} K^{-1}, respectively. Jokinen and Kukkonen (1999a,b) have given the thermal conductivity values for the upper, the middle, the lower crust and the mantle as 3.5, 3.0, 2.5 and 4.0 W m^{-1} K^{-1}. Turcotte and Schubert (2002) suggested that the thermal conductivity of the basalts and the granites is 1.3–2.9 and 2.4–3.8 W m^{-1} K^{-1}, respectively. He et al. (2009) suggested the use of thermal conductivity values of 2.9, 2.8, 2.5 and 3.0 W m^{-1} K^{-1} for the upper, the middle, the lower crust and the upper mantle, respectively. Thermal conductivity varies only little in typical lithospheric rock types and is usually between values of 2 and 4 W m^{-1} K^{-1}, but conductivity is also dependent on the temperature.

According to Verzhbitsky (2002) and Galushkin et al. (2006), the thermal conductivity values of the upper, the middle and the lower crust for the eastern Black Sea basin are 1.2, 1.3 and 1.6 W m^{-1} K^{-1}, respectively. Galushkin et al. (2007) used the thermal conductivity in the region of the Shatsky and Andrusov rises as 2.72, 2.72 and 1.88 W m^{-1} K^{-1} for the granitic and basaltic layers, respectively, as suggested by Baer (1981).

Heat production values depend variably on the lithological and geochemical characteristics of the rock formations. Whereas depleted mantle shows low heat production values, upper crustal rocks have high heat production values (Moisio and Kaikkonen, 2006). In the upper part of the crust, the heat production is $A(z) = A_0 \exp(-z/D)$, z is the depth in km, A_0 is the surface heat

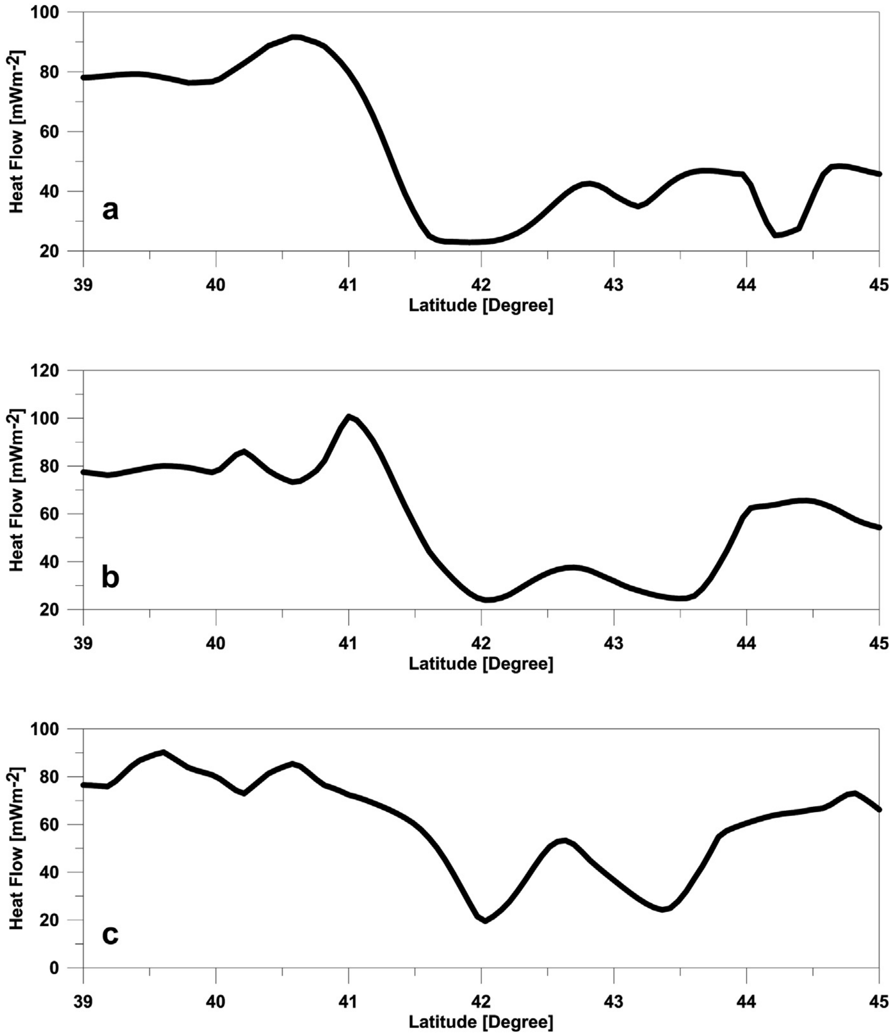

Figure 3. Surface heat flow density values along the 38° (a), 39° (b) and 40° (c) latitudes. Profile locations are shown in Fig. 2b.

production in μW m^{-3} and D represents thickness of the crustal layer in km which is radioactively enriched (Pollack and Chapman, 1977). Factor D is derived from the surface and the reduced heat flow and generally ranges from 5 to 15 km with an average of 10 km (Morgan and Sass, 1984; Pasquale, 1987).

He et al. (2009) used the heat production values for the upper, the middle, the lower crusts and the mantle as 1.10, 0.83, 0.37, and 0.24 μW m^{-3}, respectively. In addition, Jokinen and Kukkonen (1999a) used the heat production values for the upper, the middle and the lower crust and the lithospheric mantle as 1.8, 0.6, 0.2 and 0.002 μW m^{-3}, respectively.

In the eastern Black Sea basin, Verzhbitsky (2002) and Galushkin et al. (2006) suggested to use of heat production values for the upper, the middle, the lower crust as 0.85, 1.15 and 1.55 μW m^{-3}, respectively. In the region of the Shatsky and Andrusov rises, the heat production values are used as 1.26, 0.71, 0.21 and 0.004 μW m^{-3} for the granitic and basaltic layers, respectively, as suggested by Baer (1981) and Galushkin et al. (2007).

5. Results and discussion

5.1. Temperature and heat flow density in the lithosphere

In this study, Moho temperature, Curie point depth and thickness of the eastern Black Sea basin and the eastern Pontides orogenic belt lithosphere are obtained from thermal modeling with constraints from surface heat flow measurements. Geothermal modeling results indicate that the Moho temperature values (Fig. 4a) vary between 367 °C in the eastern Black Sea basin, where the low heat flow values are observed, and 978 °C in the eastern Pontides orogenic belt where the high heat flow values exist by using surface heat flow ranging from 14.5 to 100.9 mW m^{-2}.

In the eastern Black Sea basin the Moho depth is about 21.5 km and increasing to a depth of 52 km in the eastern Anatolia. In this region, both surface and mantle heat flow values are rather high with respect to the Black Sea basin. The estimated heat flow values (Fig. 4b) at the base of the crust fall in the range of 16.4–56.9 mW m^{-2}, with low heat flow values in the eastern Black Sea basin relative to the eastern Pontides orogenic belt.

Curie point depth values (Fig. 4c) beneath the study region vary from 19.8 to 76.0 km. In the continental crust, mean Curie point depth value is about 25 km which is thinner than the Moho depth. However, in the eastern Black Sea basin, the depth range of the Curie point surface values is rather large and thicker than the Moho depth.

Deep Curie point depth values are observed in oceanic areas. This is consistent with low heat flow data in these regions. Shallow Curie point depths lie in the back-arc regions, trenches and consistent with high heat flow data. Higher heat flow values occur

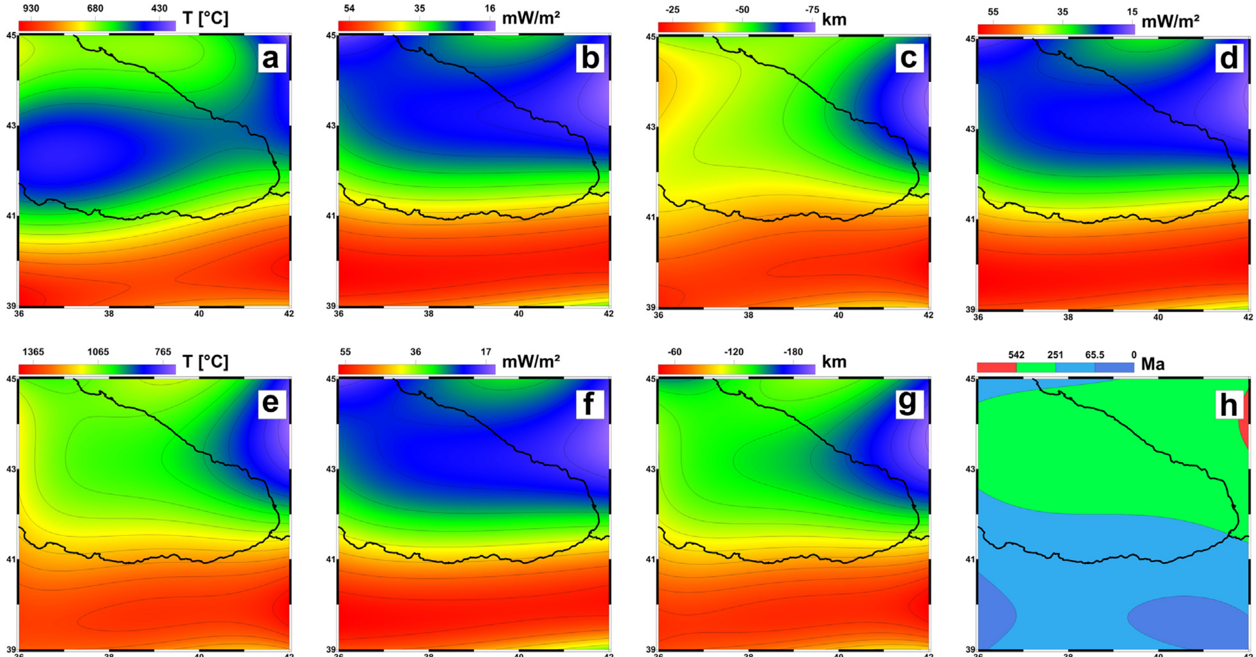

Figure 4. Map of Moho temperatures (a), Moho heat flow density (b), Curie point depth (c) and the Curie heat flow density (d) beneath the study region. The temperature (e) and heat flow density values (f) at the depth of 100 km. The lithosphere–asthenosphere base (g) and the age of the lithosphere (h) according to Parker and Oldenburg (1973) equation for the eastern Black Sea basin and eastern Pontides orogenic belt. The temperature of the asthenosphere is assumed as 1300 °C.

at the edge of the continental margin and are consistent with a sudden change of the Curie point depth. Previous studies show that the Curie point depths, which are greatly related to the geological context, are shallower than about 10 km at 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 (Yamano, 1995; Tanaka et al., 1999).

The heat flow values (Fig. 4d) at the Curie point surface for the investigated region are ranging between 14.3 mW m^{-2} in the eastern Black Sea basin and 59.4 mW m^{-2} in the continental crust. Along the southern Black Sea coast, the computed heat flow values increase markedly at around the trench within a few kilometers to the south thorough the arc region. This sudden increase might be caused mainly by thicker continental crust.

Temperature values (Fig. 4e) at the depth of 100 km change from 706.8 °C in the eastern Black Sea basin (trench) to 1432.1 °C in the arc region. Along the southern Black Sea coast, there is a sudden change in temperature values which are parallel to the coastal line. Within the continent, maximum temperature values are obtained in the arc region. Heat flow values (Fig. 4f) at the depth of 100 km range between 14.5 mW m^{-2} within the eastern Black Sea basin and 56.9 mW m^{-2} in the continent which show the trench has relatively low geothermal potential with respect to the back-arc. The reason for the lower heat flow values located in the southern Black Sea coast might be related to the input of water into the subduction zone.

According to Artemieva (2006), lithospheric geotherms in tectonically active continental regions are assumed to reach 1300 °C at a depth of 60–100 km as indicated by xenolith geotherms, whereas the lithospheric thickness in modern zones of continent–continent or continent–ocean collision (Andes, Hellenic arc, Alps) was fixed at 220 km. In this study, the lithospheric thickness, where the temperature is 1300 °C isotherm, is reached at depth range between 49.4 km in the eastern Pontides orogenic belt and 152.2 km in the Black Sea basin (Fig. 4g).

Parker and Oldenburg (1973) suggested that a relationship between the depth and the age of the lithosphere according to a ratio based on a connection between the thickness of oceanic lithosphere and the time of its cooling:

$$H = 7.8\sqrt{t} \tag{3}$$

where H is the thickness of the lithosphere (km) and t is the age of the lithosphere (Ma). In this study, the lithosphere age is determined according to the Eq. (3) from the thickness of lithosphere for the study region (Fig. 4h). The time of formation of Black Sea basin range from Paleozoic to Mesozoic, suggesting that the Black Sea is a remnant of the Paleotethys Ocean as proposed by Chorowicz and Dhont (2002). On the other hand, the age of the eastern Pontides orogenic belt corresponds to the Mesozoic and Cenozoic.

Fig. 5 shows the tectonic elements of the Black Sea basin (Alptekin et al., 1986) and the thrust faults that occur at the subduction zones (Barka and Reilinger, 1997). The southern margin of the Black Sea has not been studied in detail due to the low rate of seismic activity. However, the 1968 Bartın earthquake occurred along the thrust faults that lie parallel to the southern Black Sea coast (M = 6.8) which testifies that this margin can produce destructive earthquakes (Fig. 5).

5.2. Geodynamic implications

The origin of the eastern Black Sea basin and Pontides orogenic belt is still controversial due to lack of systematic geological, geophysical and geochemical data. It is widely accepted that the eastern Pontides orogenic belt was shaped by northward subduction of Neotethys or Paleotethys oceanic lithosphere during the late Mesozoic and the Black Sea basin opened as a back-arc basin behind the eastern Pontides magmatic arc (Şengör and Yılmaz, 1981; Okay et al., 1994; Dilek et al., 2010). However, both results obtained from this study (Fig. 6a and b) and also most recent paleomagnetic, geological, geochemical and geochronologic studies on the late Mesozoic–Cenozoic geodynamic evolution of the eastern Mediterranean region do not support this idea. The common features of back-arc basins that are associated with subduction in the relation of

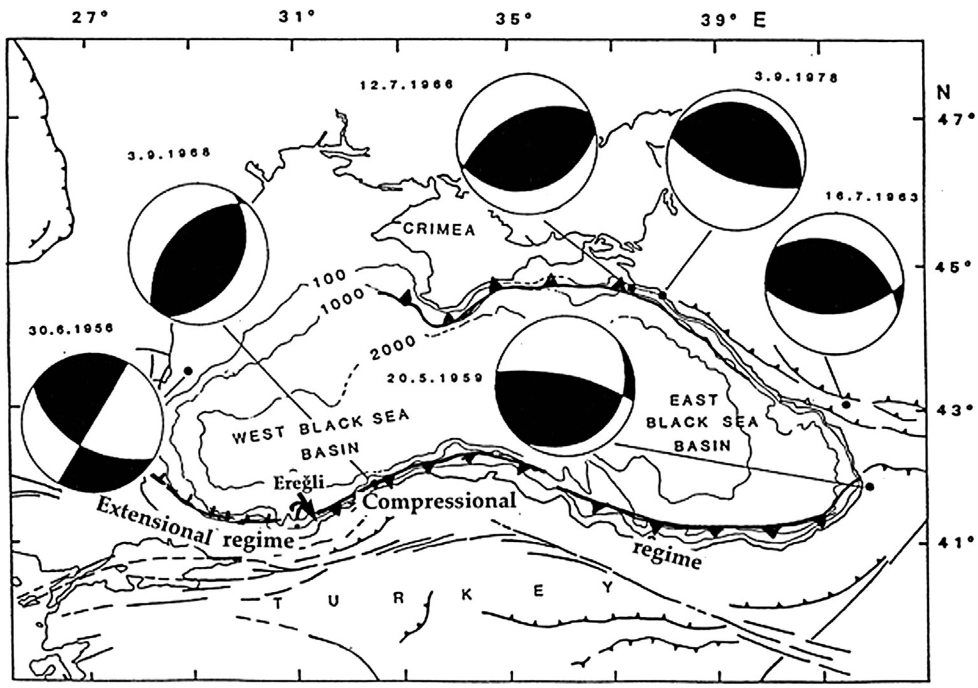

Figure 5. The main tectonic structures of the Black Sea basin (Alptekin et al., 1986) and the reverse faults occurred at the subduction zones of the Black Sea basin (Barka and Reilinger, 1997).

surface heat flow (Fig. 6a) are summarized as: (1) less than normal heat flow over the zone from trench axis to volcanic zone; (2) high, but variable, heat flow over the volcanic zone or island arc; and (3) the mean heat flow in back-arc extensional basins (Uyeda, 1977).

The paleomagnetic studies indicate that the eastern Pontides orogenic belt was situated at 23° (Van Der Voo, 1968; Lauer, 1981; Sarıbudak, 1989), 25.5 ± 4.5° (Channell et al., 1996), 20.0 ± 2.5° (Çinku et al., 2010), 26.6° (Hisarlı, 2011) north latitude in the Cretaceous. The present day location of the orogenic belt is between 39° and 42° north latitudes. This northward displacement since the Cretaceous can best be explained by southward subduction of the Tethys oceanic lithosphere.

In the eastern Pontides orogenic belt, the late Mesozoic and Cenozoic are characterized by intense igneous activity. The detailed investigations on the late Cretaceous magmatism indicate that magmatic activity started with TH–CA character during the Cenomanian–Turonian in the northern part of the magmatic arc and migrated southward. This activity shows a transition to CA–A magmatism in the southern part of the orogenic belt in the Coniacian–Campanian. Further south, the igneous activity passed into shoshonitic to ultrapotassic volcanism during Campanian–Maastrichtian (Bektaş et al., 1999; Eyuboglu, 2010). Conversely, systematic geological, geochemical and geochronological studies on the Cenozoic magmatic rocks indicated that the Cenozoic magmatism started in the late Paleocene in the far south of the magmatic arc and migrated toward north in time. According to Eyuboglu et al. (2011a,b,c,d, 2012b), the southward migration and increasing potassium content of the late Cretaceous igneous activity and also northward migration of the Cenozoic magmatism might be explained by a roll back of the Tethys oceanic lithosphere, which was subducted toward the south during the late Cretaceous, starting from Tertiary (Fig. 6b).

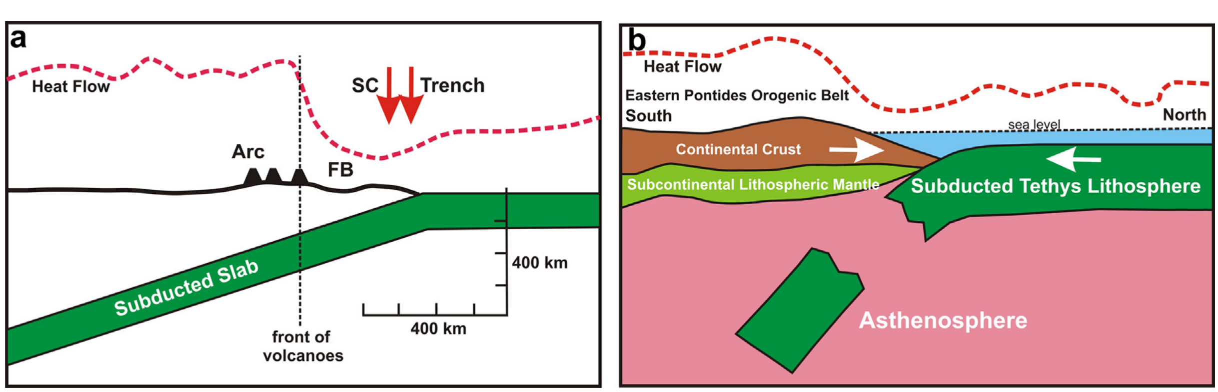

Figure 6. (a) Typical surface heat flow variations near subduction zones (Uyeda, 1977). General heat flow values are expected to decrease on the subducting plate approaching the trench. FB: fore-arc basin; SC: subduction complex. (b) The late Cenozoic geodynamic evolution model for the eastern Pontides orogenic belt (after Eyuboglu et al., 2012b) and also heat flow values obtained from this study.

6. Conclusions

The following conclusions can be drawn from the present study:

(1) In the eastern Pontides orogenic belt, the highest heat flow values (105 mW m^{-2}) associated with extensive Neogene and Quaternary volcanic activity are observed.
(2) The lowest heat flow values (12 mW m^{-2}) are seen in central part of the eastern Black Sea basin.
(3) The Moho temperature increases from 367 °C in the eastern Black Sea basin to 978 °C in the eastern Pontides orogenic belt.
(4) At the base of the crust, computed heat flow values are range of 16.4–56.9 mW m^{-2}. High heat flow values might be related to melting of the lithospheric mantle caused by upwelling of the asthenosphere.
(5) Curie point depth values for the study region change between 19.9 and 76.1 km. Mean Curie depth value is about 25 km which is thinner than the Moho depth value in the continental areas.
(6) The heat flow values at the Curie point surface increase markedly from 14.3 mW m^{-2} in the trench region to 59.4 mW m^{-2} in the arc region, which is caused mainly by a thicker crust.
(7) At the depth of 100 km, temperature values change from 706.8 °C in the trench region to 1432.1 °C in the arc region. The cause of the lower temperature values located in the southern Black Sea coast might be related to input of water into subduction zone.
(8) The thermal lithospheric thickness varies between 49.4 and 202.1 km, decreasing southward from the Greater Caucasus to the eastern Pontides orogenic belt. In the eastern Black Sea basin, the maximum thermal lithospheric thickness reaches 152.2 km where the temperature value is 1300 °C.
(9) Along the southern Black Sea coast, the trench region has relatively low geothermal potential with respect to the arc and back-arc region.
(10) The time of formation of Black Sea basin occurred from Paleozoic to Mesozoic, showing the Black Sea is a remnant of the Paleotethys Ocean.
(11) The numerical studies give indication of southward subduction model for the origin of the Pontides during late Mesozoic–Cenozoic.

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