PERGAMON

Journal of Geodynamics 33 (2002) 201–218

JOURNAL OF GEODYNAMICS

www.elsevier.com/locate/jgeodyn

Eastern Pontides and Black Sea: gravity inversion, crustal structure, isostasy and geodynamics

H. Çavşak[a], W.R. Jacoby[b,*], A. Şeren[a]

[a]*Jeofizik Mühendisligi Bölümü, Karadeniz Teknik Universitesi, Trabzon, Turkey*
[b]*Institut fw. Geowissesnschaften, Johannes Gutenberg-Universität Mainz, Saarstr. 21, D-55099 Mainz, Germany*

Received 18 October 1999; received in revised form 9 July 2001; accepted 9 July 2001

Abstract

Lacking detailed seismic data on the crustal structure of Anatolia and the transition to the Black Sea, it is attempted to invert gravity for crust-mantle structure with constraints from the limited a priori information available, as average continental and oceanic crust, local topography, an isostatic model, published marine seismic data and a tentative "fix point" from recent seismological receiver functions near the Black Sea coast. An initial 2D-model for a north–south profile along about 40°E longitude is constructed and adjusted to fit the Bouguer anomaly taken from a published Turkish map and results from satellite radar altimetry. Isostasy, seismic data and gravity inversion concur in suggesting that the Moho under the eastern Black Sea is about 25 km deep with 15 $\pm$ 1 km thick sediments (4–7 km unconsolidated). Under eastern Anatolia where the average elevation is about 2 km, the Moho comes out to be about 55 $\pm$ 5 km deep. No constraints exist for the upper-lower crust transition. The gravity fitting model is not far from an isostatic mass balance, but the high topography of $\sim$2 km seems slightly overcompensated with the Moho $\sim$3 km deeper than predicted by the isostatic model; a slightly increased mantle density beneath eastern Anatolia would restore the isostatic balance without contradicting gravity. We favour such a vertical balance over a regional or lateral mass balance inherited from the initial opening of the Black Sea basin.

1. Introduction

Understanding tectonics and geodynamics of eastern Anatolia and the adjacent Black Sea requires some knowledge of crustal structure and isostatic setting, presently nearly unknown. It is, therefore, attempted to collect the available data and to use gravity inversion with a priori information to constrain the structure of the transition from sea to land along a representative profile.

* Corresponding author. Fax: +49-613-139-4769.
E-mail address: jacoby@mail.uni-mainz.de (W.R. Jacoby).

0264-3707/02/$ - see front matter
PII: S0264-3707(01)00064-3

Relevant seismic information exists about the crust of the eastern Black Sea Basin (Belousov et al., 1988), however, scarce in the area of interest. Several wells have been drilled for exploration purposes into the upper sedimentary layers. Heat flow has been measured at many points, but it does not constrain crustal structure.

Very little is known from eastern Anatolia, but there is a tentative fixpoint near the coast at Trabzon from a receiver function study (Cakir et al., 2000) and from low resolution Pn tomography for the Turkish–Iranian plateau (Hearn and Ni, 1994). Gravity, though non-unique, can be used to test density models and to explore various hypotheses.

An appropriate method is inversion, starting with the linear aspect of density adjustment and then exploiting the possibilities of general non-linear inversion. Two-dimensional models suffice in view of the scarcity of constraining data and it is appropriate for a section at a right angle to the Black Sea coast; a N–S profile is chosen for gravity modelling and inversion along 40°E longitude. No similar work is known to the authors with the exception of Spadini et al. (1996) who applied dynamic modeling of rifting and did some forward gravity modelling.

First, a brief review of geology and geophysics of the region, as far as known, summarizes the data on topography and bathymetry, geology and gravity anomalies for Turkey and the Black Sea for which relevant seismic refraction data are available. Present knowledge of general crustal properties are taken into account. The method and model assumptions with the constraining data are then presented, and the modelling results are analyzed in terms of two possible isostatic adjustments to earlier and ongoing dynamic processes. Spadini et al. (1996) discussed a regional isostatic model based on the idea of backarc opening and subsequent evolution of the western and eastern basins of the Black Sea. Their results will be compared to ours with the aim to investigate the isostatic situation in the geodynamic setting of adjacent convergent and divergent structures. They are discussed in the context of the evolution of the north Anatolian–Black Sea continental margin.

2. Geological setting

N.I. Andrusov, a Russian geologist, was the first to explore the Black Sea in two expeditions onboard navy ships in 1890 and 1891, collecting samples from shallow and deep waters of the whole Black Sea. The first modern maps of bottom relief were compiled by Goncharov et al. (1972). Some authors imagined the Black Sea to be a Neogene or even Quaternary graben-like gap within the continental crust (Andrusov, 1893; Dobrynin, 1922) or a current geosyncline (Obruchev, 1926; Nalivkin, 1928; Arkhangelsky and Strakhov, 1938). The probable absence of a granite layer was explained by a primarily oceanic origin (Muratov, 1955; Milanovsky, 1963, 1965). The thickness of the sediments suggested Paleozoic or Upper Pre-Cambrian age. Belousov considered the Black Sea a result of basification (oceanization) of continental crust. Modelling heat flow led Golmshtok et al. (1992) to conclude that the basin is Jurassic in age. In mid Jurassic and Late Cretaceous, the basin lay north of major volcanic arcs exposed in the eastern Pontides. Kropotkin (1967) was the first to consider the Black Sea to have originated by horizontal extension.

The Black Sea which has a uniform depth of 2.2 km is today considered the result of Mesozoic to Cenozoic backarc spreading in several phases. There are no clear marine magnetic anomalies; indistinct original anomalies and deep burial may be the reason. The Black Sea frame is tectonically heterogeneous (Fig. 1). Regional structures have different crustal ages and geological nature (Belousov et al., 1988). In the east, the Black Sea is limited by the Kolkhids depression. The assemblage of surrounding Alpidic orogens includes the Pontides in the south and the Caucasus and its western extension, the Shatsky Ridge and the Krimean mountains, in the north. Tethys closure with northward subduction and collision of various continental blocks or terranes and related backarc opening of marginal seas formed the region (Ross et al., 1970, 1974a,b; Letouzey et al., 1977; Zonenshain and LePichon, 1986; Barka and Kadinsky-Cade, 1988; Yilmaz, 1993; Okay et al., 1994; Bektas et al., 1995). The transition from continental NE Anatolia to the oceanic Black Sea is marked by Mesozoic volcanics and sediments with evidence for convergence (Ross et al., 1970; Yilmaz, 1993). A significant part of the platform and folded structures surrounding the Black Sea continue also on its shelf to the continental slope into the deep sea basin (Fig. 1).

The western basin is thought to have formed in an earlier Cretaceous (Cenomanian) phase by rifting in the stable Moesian platform and subsequent seafloor spreading. The eastern basin,

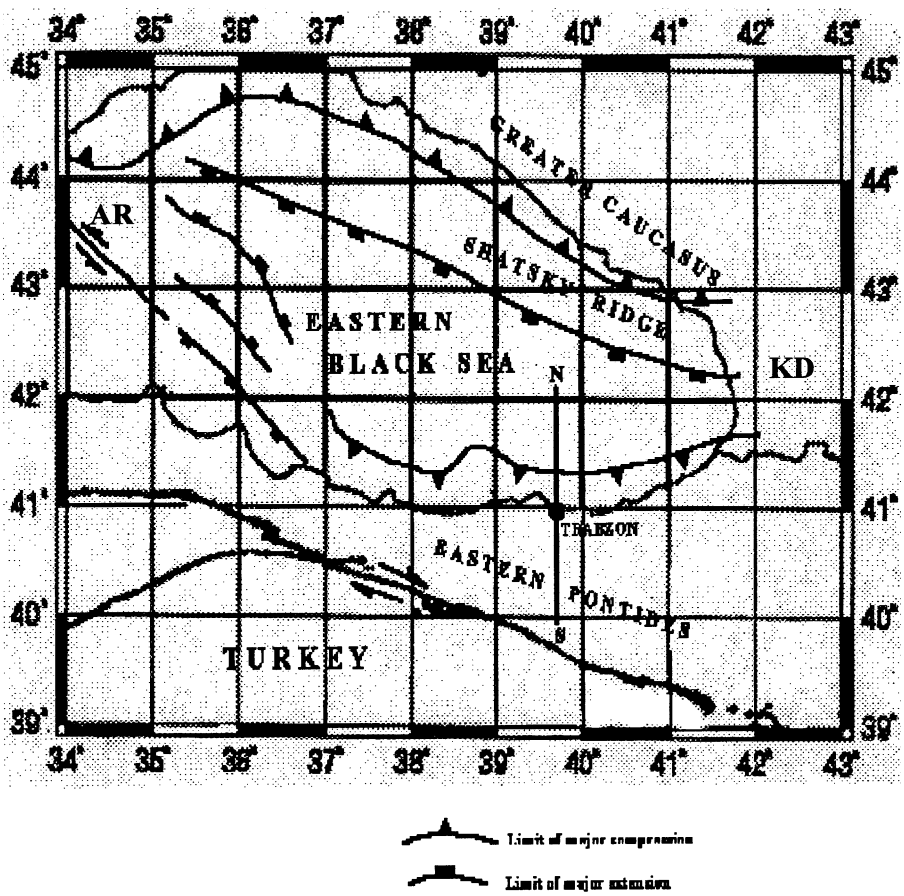

Fig. 1. Sketch map of eastern Black Sea and Anatolia, with modelled profile shown as thick N–S line. The major tectonic elements are shown: the Mesozoic-Cenozoic fold belts eastern Pontides in the south (with the north Anatolian Fault) and Greater Caucasus in the north. The eastern Black Sea basin is bounded in the north and southwest by rifted margins, in the south by a convergent margin. AR: Arkhangelsky rise, KD: Kolkhids depression.

studied here, is believed to have had a more complex history which cannot be fully resolved; there are indications for extension and subsidence since early Jurassic; the basin formed by rifting and spreading during the late Paleocene, complete by Middle Eocene. The two basins are separated by a basement high (Arkhangelsky Rise) which is totally covered by younger sediments and may have rotated clockwise from the Shatsky Ridge of the NE Black Sea shelf.

The basins are filled with thick sediments of post-spreading ages. The complex eastern basin contains at its margins Jusassic to Cretaceous sequences of sediments and volcanics. While seismic structure in the western basin suggests oceanic crust covered by 15 km sediments, this is not so clear in the eastern basin. The suggested 13 km sediments under 2 km water would leave 10 km igneous crust if the Moho is at 25 km depth, as determined by refraction seismic profiling (Belousov et al., 1988). It has been suggested that the crust in the eastern Basin may not be entirely oceanic. However, the seismic refraction data discussed below are not so certain.

Spadini et al. (1996) assume that the two basins originated by rifting of very different lithosphere, about 200 km thick in the west and 80 km thick in the east. This would influence the depth of necking and subsequently the basin morphology and fill. Especially for the eastern basin, the original crustal thickness is very uncertain because of the complex earlier history; Russian Platform crust has a characteristic thickness of 40–45 km (Guterch et al., 1986). But in view of the preceding history, the crust may have been thinned.

3. Geophysical data

Topography and bathymetry are taken from topographic maps (1:25 000) of the state institution and from the "Bathymetric Chart of the Black Sea" (modified from Ross et al., 1974b) with contour intervals of 200 and 2000 m.

Central to this study are the gravity anomalies taken from the map published by the state institution "M.T.A. Genel Mudurlugu ve Harita Genel Mudurlugu". No detailed information is available on the field measurements, instruments and their errors. It is assumed that the main error source is the drift of the instruments. The gravity data are referred to the Potsdam base value 981 274 mGal and the international gravity system of 1967. The Bouguer density was assumed to be 2670 kg/m^3. The topographic (relief or terrain) reduction was done with the aid of the Hammer tables out to zone J. Reference height is mean sea level. A grid size of 10×10 km^2 has been applied. Contour interval of the map is 5 mGal. For the Black Sea region the M.T.A. gravity map shows also Bouguer anomaly contours, but no information is available on the data quality (M.T.A., 1994). In the inversion we shall normally asume a nominal error of 10 mGal.

Also consulted is the map of gravity anomalies, based on satellite radar observations (ARK, no year). The map presents Bouguer anomalies which are obtained from sea surface topography via Free Air anomalies by "filling in" the water to rock density 2.670 kg/m^3. The ARK map uses "data from the 1969 ATLANTIS II cruise, a Russian chart supplied by Pavel Kuprin of the University of Moskow, and US plotting sheets 108N and 3408N. Land topography is from the Morskoi Atlas, Tom 1, Navigationne-Geographicheski I Izdanie Morskogogeneralnogo Shtaba. The map was contoured by Elazar Uchupi of the Woods Hole Oceanographic Institution".

The two maps (M.T.A. and ARK) have some large discrepancies. Especially in the neighbourhood of the Black Sea shore the discrepancies necessitated rather arbitrary merging (by

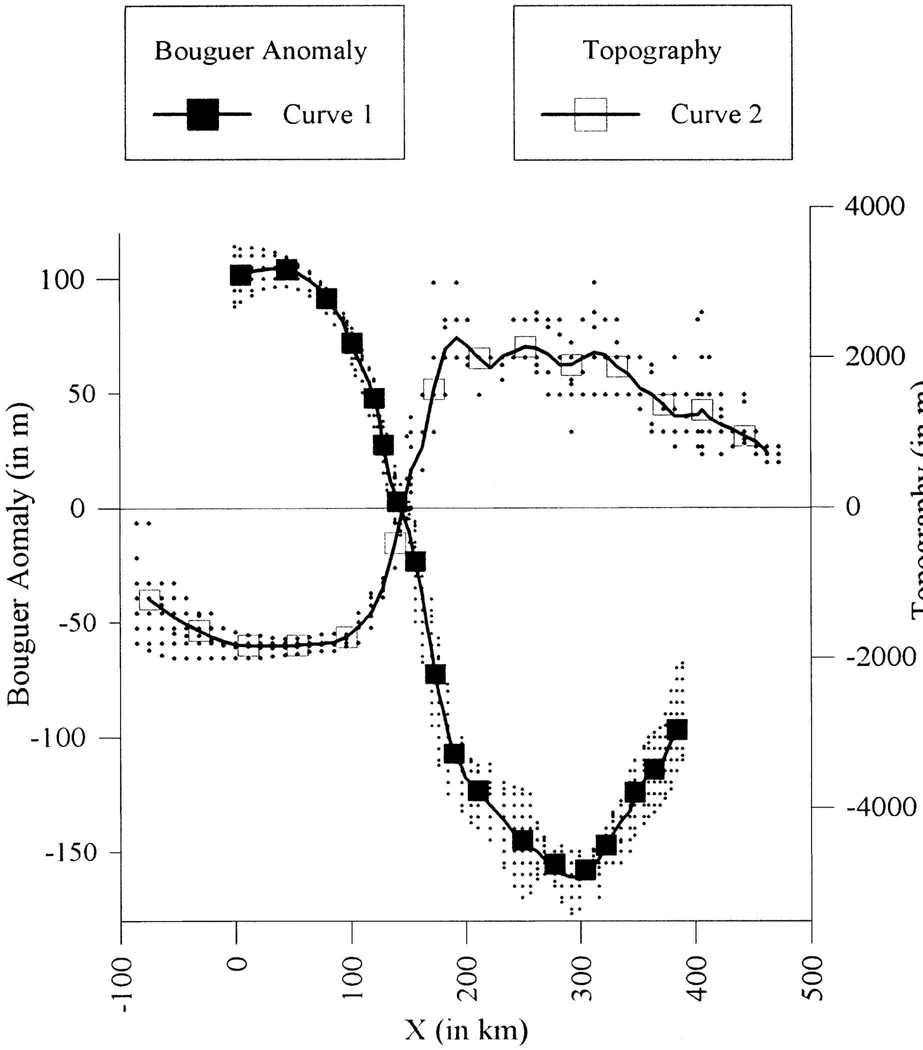

Fig. 2. Bouguer-anomaly and topography/bathymetry profile from the Black Sea to eastern Anatolia, following east longitude 40.125°, from 38.75°N to 42.25°N (see Fig. 1); averages of a set of points lying within overlapping rectangles are assigned to the center points. The strip is one degree wide. Data from "M.T.A. Genel Mudurlugu ve Harita Genel Mudurlugu". Systematic errors are likely to occur as demonstrated at sea by comparison with another data source (ARK, no year).

averaging) of the two profiles obtained from the maps, since there is no way of deciding which is better. A test is discussed below where the same geometrical initial model was subjected to the inversion of the two different gravity data sets. It turned out that the choice of one or the other Bouguer anomaly profile was not critical for the results.

The seismic receiver functions obtained for station Trabzon by Cakir et al. (2000) indicate Moho depths varying from 32 to 40 km (depending on the direction of the waves), dipping to the south, as one would expect. A low velocity lower crust below remarkably high velocities may be indicated. This is the only recent seismic control on the land side (near the coast) along the profile. A Moho depth of 36 km can be crudely estimated from the Pn tomographic study by Hearn and Ni (1994) and the station residual at Trabzon of about −1.3 s, relative to a mean Moho depth of 41 km, with mantle and crust velocities of 8.0 and 6.3 km/s, respectively (see also Kadinsky-Cade et al., 1981). For mantle structure tomography, results by Spakman (Wortel and Spakman, 2000) are consulted (see below).

Information on Black Sea structure (sediments and crust) are reflection and refraction (DSS) seismic studies and also gravimetric interpretations (Ross et al., 1970; 1974a; Belousov et al., 1988; Bogolepov et al., 1990; Orovetskii, 1999; p. 186). A recent study of seismic surface waves (Sayil and Osmansahin, in press) crossing the Black Sea gives a rough indication of a thick crust. A combination of the various seismic data for the eastern basin gives a 25 km deep Moho and possibly a relatively thick igneous oceanic crust under at least 13 km of post-rift sediments near basin centre (coordinates $x = 0$–50 km, Figs. 2–5).

Various velocity–density relationships had been assumed by different authors to convert seismic models to density models (e.g. Gardner et al., 1974; Matthews, 1939), but these are not critical to the inversion which independently determines the best fitting densities. There is no unique velocity–density relationship, only trends or "systematics".

The profile investigated crosses the Black Sea coast and follows a meridian (East longitude 40.125°). The profile is about 400 km long, starting at 42.25°N and ending at 38.75°N. In order to

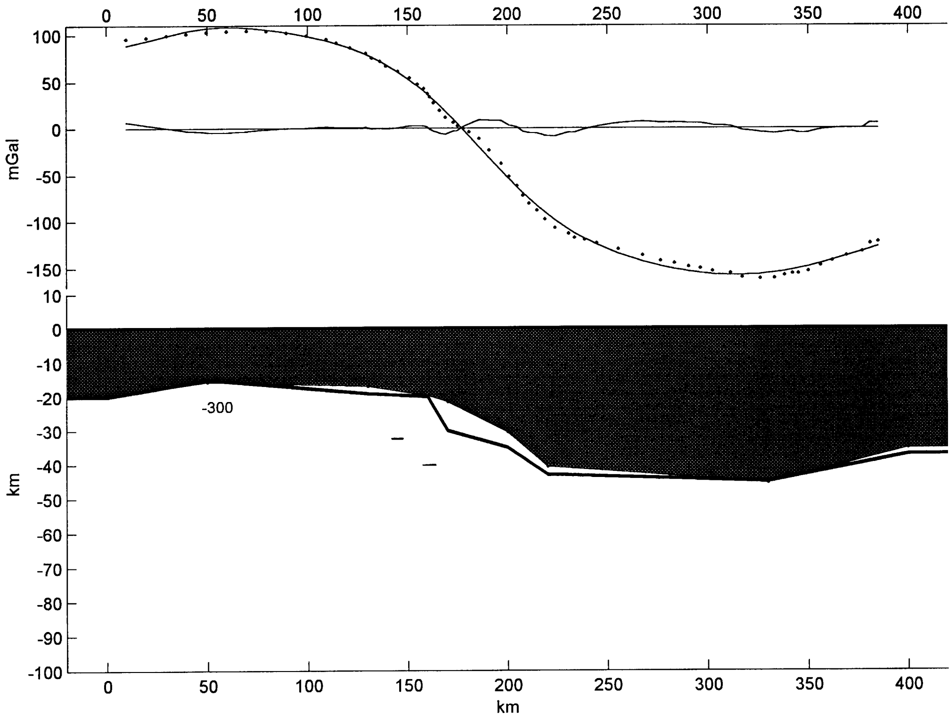

Fig. 3. Comparison of two initial models with homogeneous crust; first model (Trabzon) is shown by a line (Moho), first model (Mainz)in grey shades (initial assumptions dashed); numbers are computed density contrasts in kg/m^3. Also shown for the Mainz model are the "observed" and calculated Bouguer anomaly values (points and line, respectively) and the residuals.

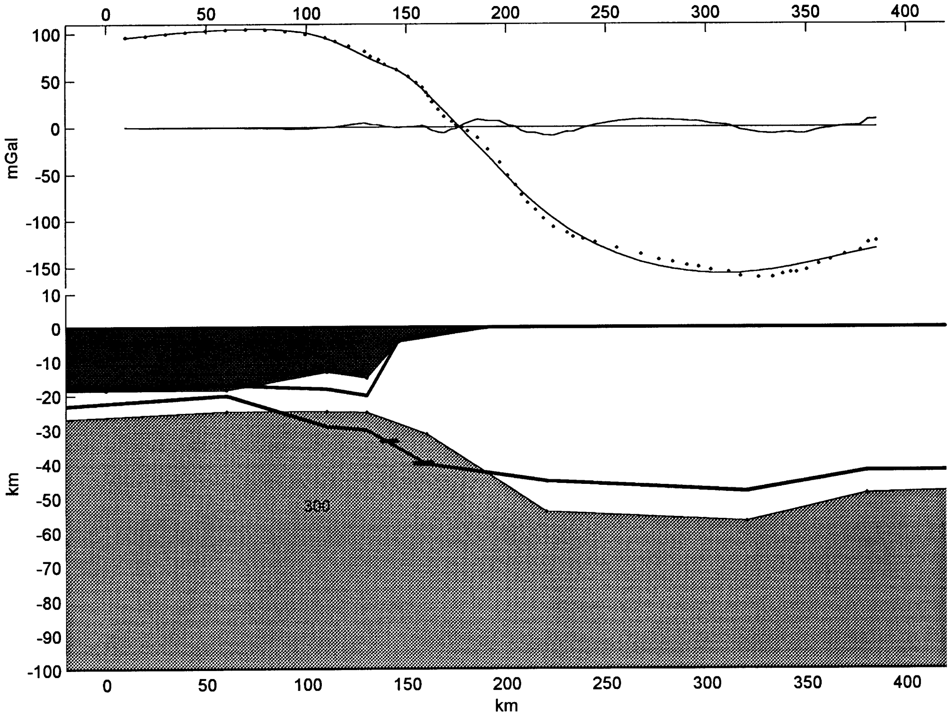

Fig. 4. Model with Black Sea sediments included; the initial assumptions were the same as for Fig. 3, except for the additional sediment body. Lines: initial model; shaded: final model.

suppress local scatter a smoothed mean profile was derived from a strip of topography (bathymetry) and Bouguer anomaly by averaging a set of points lying within overlapping rectangles and to assign the averages to the center points. The strip is one degree wide bounded by 39.625° and 40.625°E (Fig. 1).

The accuracy of the profile data used in inversion is not easy to assess. Little or no information about the original data errors is available. The standard deviations of the averaged point values may have little meaning for the task at hand because we are only interested in large scale features. A more serious question is that of systematic errors which are likely to occur and may reach values of tens of milligals. This is especially demonstrated at sea by comparison of the two data sources. We shall discuss this problem in connection with the inversion. Some inversions are repeated with different assumptions of the errors.

4. Method

Lacking detailed seismic information one can set limits on possible crustal thicknesses and their uncertainties with the aid of gravity modeling, aided by isostatic arguments. Inversion of the gravity anomalies is used as a tool. Since a priori information, always needed, is limited, general knowledge of continental and oceanic structure and its variability and geological considerations are important. NE Anatolia is a mountainous region with high average elevation, and the Black

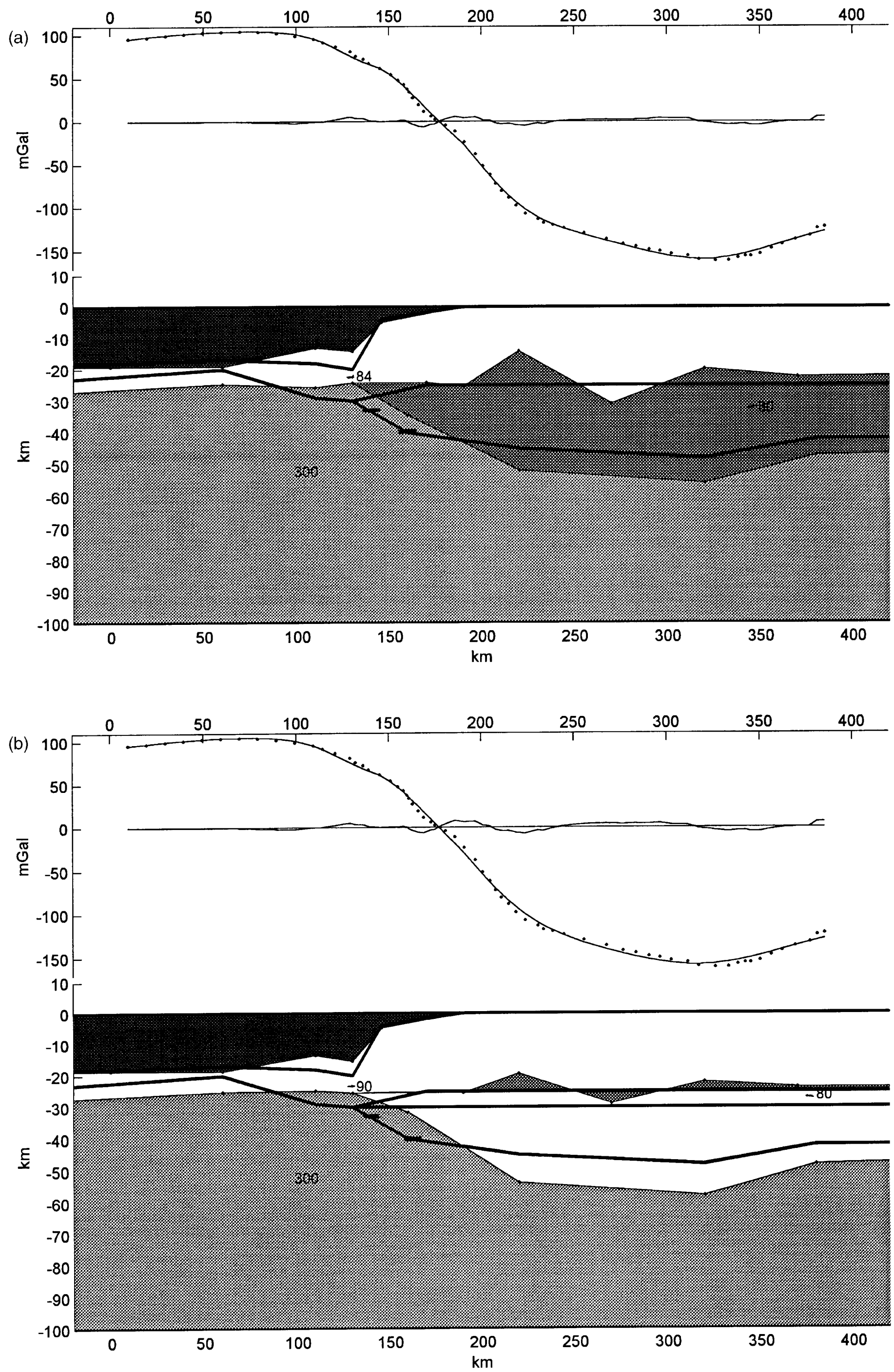

Fig. 5. Model with undulating middle-lower crust boundary, (a) density decrease with depth assumed, (b) same, but with different initial assumptions, (c) density increase assumed. Lines: initial models; shaded: final models.

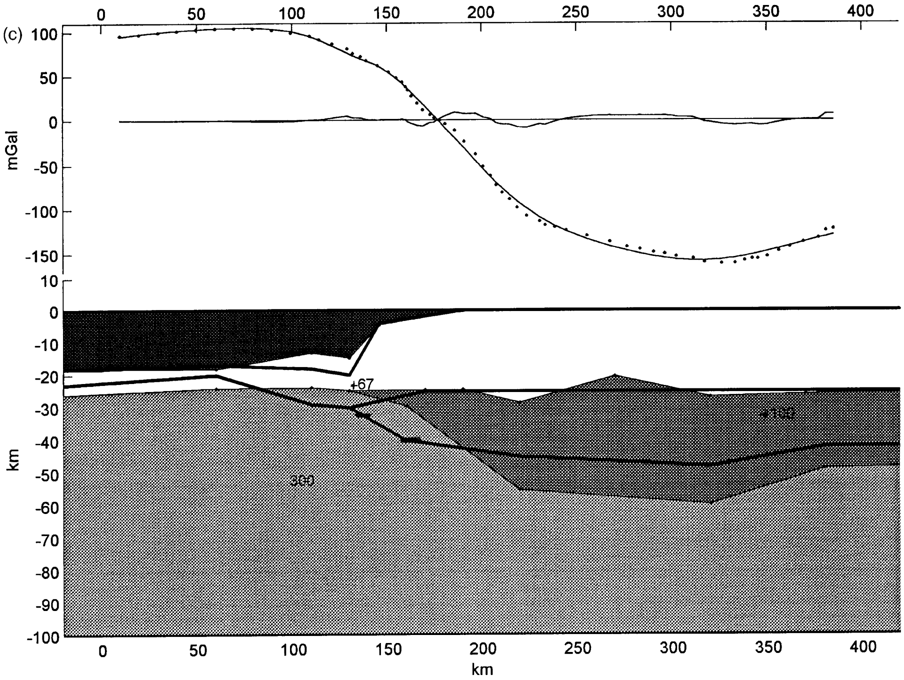

Fig. 5. (*continued*)

Sea bottom is loaded with thick sediments. Tectonic (horizontal) loading is presently governed by the north Anatolian transform fault, but at least since the Mesozoic convergence was dominant including northward subduction of Tethys oceanic lithosphere by which the Black Sea was formed as a marginal basin.

Quantitative interpretation and inversion, first of all, requires foreward routines for computing gravity effects at observation points for assumed geometrical and density models. The program used has many options, but here only 2D models were considered (at right angle to the coast) and the forward calculation of the gravity effects was done by the Talwani et al. (1959) method. The poor background knowledge does not warrant the more complicated 3D approach. The observation points are assumed at the topographic (or sea) surface.

Least squares solution of the linear problem, i.e. for the density contrasts of geometrically predefined model bodies can be used as a start (Jacoby, 1966; Cavsak, 1992). The general inversion problem is, however, non-linear. Traditionally this has been approached by "trial and error" procedures which, however, are inadequate for exploring the whole model space.

The procedure followed here is non-linear iterative inversion where the geometrical aspects of the modelling are stepwise linearized and the solution is obtained by iteration; the program package INVERT is used written by Smilde (1998). In a starting model one must specify both the a priori geometrical and density assumptions (parameters, e.g. coordinates) and, equally important, the error limits of the parameters. In the Bayesian approach applied, the parameters, including their uncertainties, describing the a priori model are considered equivalent to the data or observations. The a priori information may be from geophysical and geological results or models, e.g. seismic, or if none is available, simply guesses on the basis of comparison and experience. INVERT allows changing, and experimenting with, relative weighting of the gravity observations and the model parameters.

In INVERT the linearization is done by numerically differentating the field effects (at all observation points) with respect to all model parameters, i.e. by computing the effects at the initial values and closely neighbouring values of the parameters (under certain circumstances this may cause errors and lead the iteration astray requiring caution by the user). The linearized equations are the basis of the normal equations to be solved for the parameter adjustments. Then new residuals are computed and new parameter adjustments are computed the same way. The program allows to assume any norm, usually norm 2 (least-squares) is taken (often a final iteration with norm 1, minimizing the sum of the absolute values of the residuals, will render the most stable results).

By this procedure gravity and initial parameter values are fitted within their error limits for densities, depths and locations. It requires good judgement of the user to decide whether a result is acceptable. If the uncertainties of the initial assumptions are highly uncertain themselves, this is often very subjective, and the hope that the mathematical inversion procedure guarantees reliable objective results may fail, there is still enough subjective freedom to the limits of the solutions.

5. Model assumptions and results

The density model is defined by discontinuous-contrast boundaries. Reality is more complex, more or less gradual; it is approximated by discrete 2D bodies described by closed polygonal areas with straight-line elements (“Talwani method”). Some bodies are defined by a layer boundary with topography approximated as a polygon which is closed at some distance through the horizontal surface at sea level (having no effect on the gravity variation). This is justified as, in the Bouguer reduction, the effect of mass between the physical surface and sea level is computationally removed, if the assumed Bouguer density of 2670 kg/m^3 is correctly assumed; otherwise the residuals (observed minus computed) will contain unreduced effects of “topographic mass”; similarly the effects of sea water vs. crustal rock or marine sediments are removed computationally with the density contrast of 1640 kg/m^3. These assumptions can be checked to some extent in the inversion.

We proceed from simple to more complicated models. In the most simple case only sea water, crust and mantle are distinguished, there are two free, adjustable density contrast boundaries: crust-mantle and water-crust. In the next step, we assume three layers: marine sediments, crust (homogeneous in density) and mantle. In the next step, an intracrustal boundary between upper and lower crust is assumed, to fit intermediate wavelength gravity anomalies (<100 km wavelength, <5 mGal amplitude) by allowing undulations to adjust in the inversion. The receiver functions by Cakir et al. (2000) imply a low-density layer around 25 km depth, but its lateral extent is unknown. It is, therefore, uncertain what kind of density change with depth should be assumed; therefore, both increase and decrease are tested; note, that if it were a decrease, this is superimposed by the laterally averaged density increase with depth so that no absolute decrease with depth is implied; the horizontal average of all layers is “invisible” in gravity. Absolute densities can be assumed only from seismic velocities or “experience”. But for land there is no detailed seismic information.

If two bodies (e.g. crust and water) are defined geometrically as separate, non-penetrative bodies, their density contrasts are to be referenced against the same background; here the mantle is taken as reference. Or, if the water body is inserted into the crustal body with the contrast against the crust, its density contrast is relative to the assumed Bouguer density, because the sea had already been "filled up" in the reduction. Inversion rendered this density as <20 kg/m^3 different from standard Bouguer density (2670 kg/m^3). The land topographic mass also is close to the standard. All other separate model bodies within bigger ones are also referenced to the latter.

Inversion becomes unstable if the individual gravity effects, computed for the various bodies, correlate spatially among each other (or with a neglected body effect); this includes the case of "compensation", i.e. negative correlation. Wrong densities may result. Only a priori knowledge, with small error limits or fixed density contrasts would get us around this problem.

Forty-one kilometer average crustal thickness (as such not resolved by gravity) of the tomographic Pn study of Hearn and Ni (1994) is taken as reference, and Airy isostasy is assumed to estimate variations. Cakir et al. (2000) estimated the crustal thickness near the Black Sea coast close to Trabzon to vary from 32 to 40 km, increasing from N to S, assumed here as a "fix point" with ± 2 km error near Trabzon ($x \approx 15$ km in Figs. 2 to 5). The Moho depth would increase from the coast to the high mountains (2.2 km, 2670 kg/m^3) to 52 km, taken as the a priori information with ± 10 km uncertainty. The assumed density contrast at the Moho of 300 $\pm$ 50 kg/m^3 is relatively low, but it is appropriate in other parts of the world (Jacoby, 1973). A tentative Moho map (based on gravity and isostasy) by Makris and Stobbe (1984) reproduced by Schindler and Pfister (1997) is consulted, showing a minimum Moho depth of 24 km in the eastern basin and a maximum of 48 km along our profile in East Anatolia. High attenuation of Sn propagation in northern East Anatolia, but efficient propagation across the Black Sea (Kadinsky-Cade et al., 1981) is also considered, but has a too low resolution to constrain the models.

For the Black Sea area, two different initial assumptions are made. One is a similar isostatic model, but with three density contrasts versus average continental crust: water (maximum depth 2.2 km, -1640 kg/m^3), sediments (x km thick, -200 kg/m^3), basaltic crust (7 km, 0 kg/m^3, i.e. same density as the continental crust) and mantle [(41–2.2–7–x) km, $+300$ kg/m^3] contribute to the mass balance with average continent, giving an initial (a-priori) $x = 12.2$ km sediments and a maximum Moho depth of about 21 km in the eastern basin. The second is taking the E–W refraction line of Belousov et al. (1988) with 13 km sediments, 10 km igneous crust and 25 km Moho depth in the mid eastern basin. The uncertainties are difficult to estimate; for depths we assume values of $\pm$(3–10) km and for densities $\pm$(50–100) kg/m^3.

With an all homogeneous crust and -300 ± 50 kg/m^3 assumed for crust vs. mantle density contrast with intial Moho depths similar to those of the Makris and Stobbe (1984) map and no fix near Trabzon, gravity can be fitted by rather small depth changes (increase of <30 km from sea to mountains) and a density contrast of -293 kg/m^3; the Moho near the coast is shallower (only 18 km deep) than estimated with receiver functions (model not shown). If, on the other hand, the initial model includes the receiver function estimates near Trabzon (32–40 km, ± 2 km), inversion gives -339 kg/m^3 and a Moho depth ranging between 25 and 55 km (Fig. 3). There is no conflict with the marine seismic results and the continental estimates. But neglecting the marine sediments is not acceptable, and the model is rejected inspite of a good gravity fit of better than ± 0.5 mGal.

The next step introduces the marine sediment layer with the Moho "fix points" near Trabzon, 21 $\pm$ 5 km in the deepest Black Sea basin and 52 $\pm$ 10 km under the highest topography (see above). Lacking detailed knowledge on land, the East Anatolia crust is assumed homogeneous and no lateral density contrast assumed versus the igneous oceanic crust. For the densities used above in the initial isostatic estimates an uncertainty of ± 50 kg/m^3 is mostly assumed for the inversion. The solution (Fig. 4) thickens the marine sediments to 20 km and deepens the oceanic Moho to nearly 30 km with a minimum depth of 26 km, but shifted from the deepest basin towards the Anatolian coast. The uniform continental crust extends now to about 60 km depth between the topographic maximum and the Bouguer anomaly minimum, a region of nearly 150 km width. The computed density contrasts are now only about -160 kg/m^3 for the sediments and -399 kg/m^3 for crust vs. mantle. (the above error limits are like standard errors, hence can be exceeded, if strongly enough required by thus achieving the total least-squares solution). Assuming 2900 kg/m^3 to be a representative average crustal density, the corresponding densities of sediments and mantle would be 2740 and 3300 kg/m^3, respectively. These values call for some comments. The 2900 kg/m^3 average crustal density differ from the standard Bouguer value assumed for the reduction; since any average density increase with depth does not affect the gravity variations, a higher mean crustal density does not contradict that used for surface topography. The 2740 kg/m^3 sediment density is probably too high, but no attempt is made to correct it; with the same argument of average density increase with depth, it is likely that in the sediment depth range the continental average density is only $\sim$2700 kg/m^3 and the mean sediment density is only 2540 kg/m^3, or so.

The ± 0.4 mGal residual variations have medium wavelengths (100 km and less), and a possible cause of the medium wavelength residuals is an undulating upper/lower crust boundary or any undulations in the density depth increase or even some limited decrease, as may be indicated by the receiver functions of Cakir et al. (2000) with a low-velocity layer at mid-crustal depths. If its upper boundary undulates, a density decrease with depth would have similar gravity effects as lower boundary undulations with a density increase; the lower-crust vs. mantle density contrast might then be correspondingly smaller or greater, with larger or smaller Moho depth. These are speculations, but in any case no absolute density decrease with depth would be required by any model because of the "gravimetrical invisibility" of any average density increase with depth.

Several models have been tested. Inversion was applied to initial models with Moho and marine sediments; the additional body was defined by a horizontal string of points at 25 $\pm$ a km depth where a was taken as 5 or 10; the initial density contrast was set either + or $-$ 100 ± 50 kg/m^3. In principle, all these model types fit the observations and there is no appreciable difference between the two cases of the lower layer extending to the Moho or to a fixed depth. The optimum fit is found within the initial assumptions and the relative strengths of the depth and density constraints. Depending on the sign of the assumed density contrast, the undulations of the density boundary reverse sign (Fig. 5a–c). The particular numerical values are not significant because there is a trade-off between density contrast and undulation amplitude. The major feature of the Bouguer anomaly along the profile is the change from land to sea, and because greater mantle density ($\Delta\rho > 0$) increases the Bouguer anomaly in the sea is counteracted by a greater density contrast of the sediments ($\Delta\rho < 0$), the inversion results show a certain trade-off between the density contrasts mantle/crust and sediment/crust. Without strong additional constraints this causes uncertainty or instability of the solution.

The undulations reverse sign when the density contrast reverses sign. It may be significant that a lower crust density decrease (corresponding to the low-velocity layer suggested by receiver functions) seems to fit the data more "easily", or give more "stable" solutions, than a density increase with depth, and the undulation and density contrast amplitudes are less sensitive to other parameter variations. This is shown by Fig. 5a and b, both for density decrease (*a*: -84 kg/m^3; *b*: -90 kg/m^3). In the first case, depth undulations between 7 and 27 km (initially 25 ± 10 km) were computed; in the second case the undulations are much more moderate, ± 5 km (same as the assumed uncertainty). With the initial assumptions for depth 25 ± 5 km, density $+100 \pm 50$ kg/m^3, $+6$ kg/m^3 and no significant undulations (25 ± 1 km) resulted (Fig. 5c), but they are, of course, of reversed sign, but this model hardly deviates from a homogeneous continental crust. These models cannot be taken as a criterion for the mid crust structure and are presented only to demonstrate the uncertainties in cases of insufficient a priori information. By no means, have all the possibilities been explored here, and inversion with little constraints can produce different classes of models, e.g. with lateral density variations of the whole crust. The Moho depth maximum under the gravity minimum near $x \sim 300$ km may be an artefact of assuming laterally homogeneous crustal layers. A low-density body (e.g. -60 kg/m^3, depth extent 4–23 km) within the crust in this region would also explain the Bouguer anomaly minimum, with the Moho rather flat at 50–53 km, in this region. The "goodness" of the gravity fit cannot be used to discriminate such models; lacking such a priori information this is not persued further.

Finally the sensitivity of the results to the two differing Bouguer anomaly profiles (see above) was tested. The models show surprisingly small differences, density contrasts differing by a few percent. The largest geometrical model changes occur near the coast, where the gravity differences are largest, with basaltic crust of only 5, instead of 7 km and larger irregularities of the sediment bottom. The models cannot be distinguished by any a priori information.

6. Discussion: the isostatic state

It is of geodynamic relevance to consider the isostatic balance of the density models best fitting gravity. Isostasy need not to be perfectly maintained, especially if only the crustal masses are taken into account, but large deviations from isostasy of extended regions as the Black Sea and East Anatolia are unlikely while the possibility of deep (mantle) compensating masses always exists. A test was conducted how large the deviation from isostasy is implied in our models. Taking one of the models (all give similar results) and comparing only the "blocks" central Black Sea and the high mountains ($h \approx 2$ km) of East Anatolia, it is found that the mass columns are not exactly balanced; the Anatolian crustal root is relatively "depressed". If the mass columns would be equal, the sums, down to the deepest level, of density differencethickness/area should be equal on both sides, but they actually differ by about 10^6 kg/m^2 (equivalent to a 1000 m water or <400 m rock column). The Moho of the gravity model is about 3 km too deep under land, the topography is "overcompensated". Spadini et al. (1996) also found the best fit (to gravity) Moho of East Anatolia to be deeper than (Airy) isostatic, and under the eastern Black Sea basin shallower and explain this as inherited from original rifting with shallow necking. The implied mass excess under the basin would lead to sinking and "pulling down" the adjacent land Moho (becoming "too deep").

However, there is another possibility of compensation. Deep, not too broad compensating masses may not be "seen" in gravity inversions. Fig. 6 is a schematic representation of the two

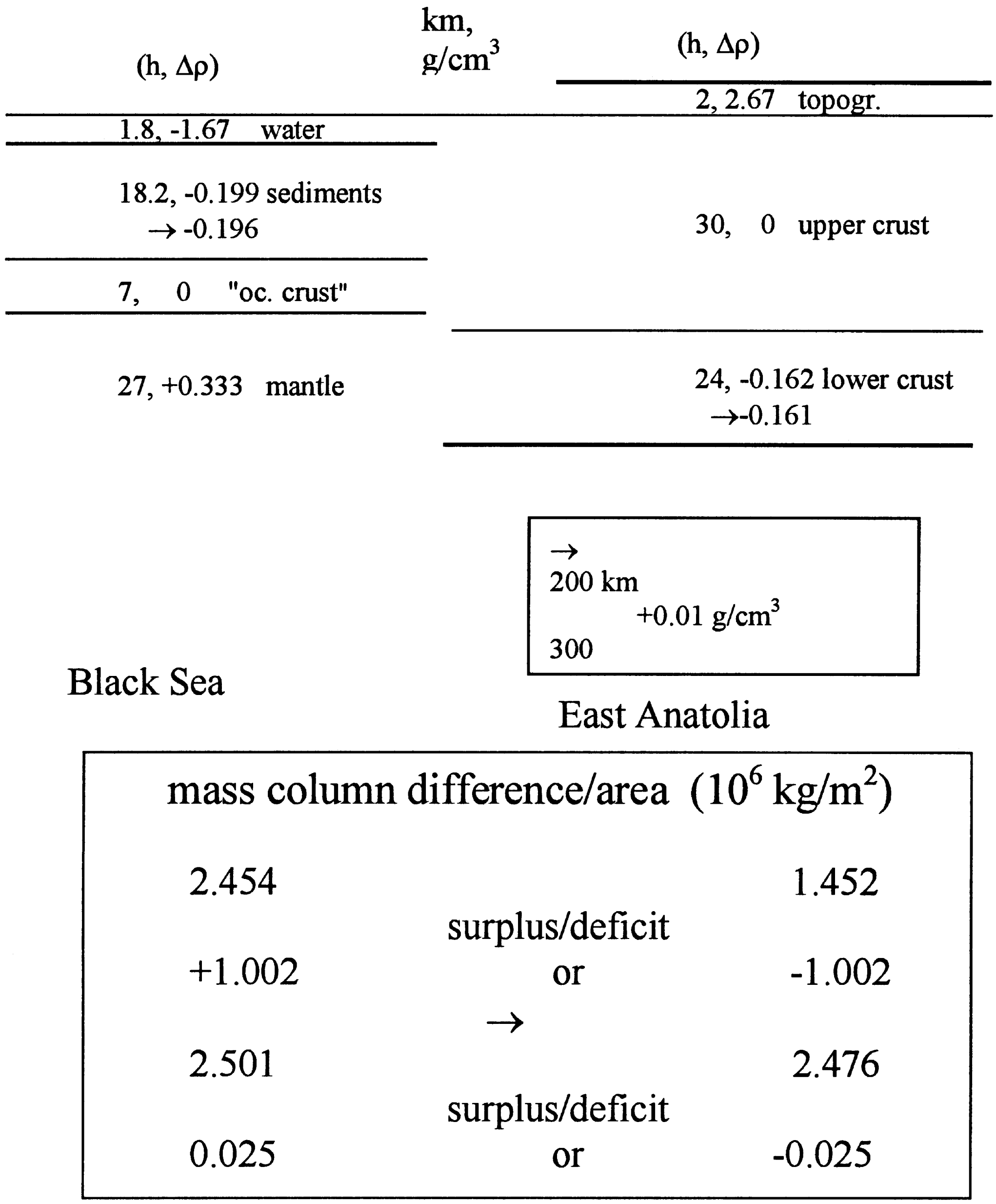

Fig. 6. Sketch of the isostatic mass balance between the Black Sea and East Anatolia, based on the gravity models without a lower crust and a "lens". One rather simple "representative model" has been taken, the isostatic analysis is for all models similar. Top: layers are depicted with their model thicknesses and density contrasts (inversion results) relative to the "crust" (7 km igneous marine and 30 km average Anatolian) as reference ($\Delta\rho = 0$); the 2D "block" below Anatolia has 100×100km dimensions (center at $x = 250$, $z = 250$ km) with a density conrast of $+10$ kg/m^3. Inversion with or without this block renders nearly identical crustal parameters. But the mass balance Black Sea versus Anatolia is affected as shown in the lower box: without the block (top lines) the mass surplus per area of the Black Sea or deficit of Anatolia amounts to 1 10^6 kg/m^2, with the block this is reduced to 2.5% (and could be reduced to zero with slightly different parameters), "restoring" the balance.

blocks giving the different layer thicknesses and lateral density contrasts (vs. the other block, reference is always the crust, but the average density and its increase with depth need not be considered); note that the model upper crust is homogeneous, i.e. the magmatic oceanic crust and the middle continental crust have no lateral density contrast. The imbalance would be removed by increasing the topographic density (between sea level and solid surface) to >3100 kg/m^3 (water deficit: -2100 kg/m^3) which is too high and would also have been seen in the inversion. If under Anatolia a deep slightly denser body is hidden (or lower density under the Black Sea undetected by gravity), mass compensation could be achieved e.g. by a body of about 100 km dimension with a density contrast of 10 kg/m^3. Such a model was tested and found compatible with both gravity and isostasy or mass balance. As also shown on Fig. 6, the added mass (marked by $\rightarrow$) between 200 and 300 km depth leads to very small changes in the inversion results (the density contrasts change by up to ~ 1 kg/m^3 and the depths by no more than a few hundred meters) and the isostatic imbalance has nearly disappeared (leaving 25 m water or 8 m rock imbalance). It is speculated that the higher-density mass below Anatolia, if it exists, may be a small relic from past plate convergence and oceanic lithosphere subduction. Wortel and Spakman (2000) report tomography results which, indeed, show a volume of relatively high p velocity at about 200 km depth below this region of eastern Anatolia.

This model is not unique; the mass imbalance estimated above (without the mantle anomaly) is within the error bounds assumed for the inversion of an idealized two-dimensional structure, thus uncertain as such. This uncertainty as such, poses the question of significance, but the "load" of 400 rock or 1000 m water is takes as significant here. It is not definite that the mass "missing" under Anatolia is where assumed, but it cannot be in shallow depths and probably not in the deep crust, since the modeling and inversion would then show it, even if with a considerable standard error. Only at greater depth its gravity effect becomes too small (and broad) to be "seen". This is a demonstation of the different sensitivity of gravity and topography (isostasy) to depth of mass anomaly.

7. Discussion and conclusions

Inspite of limited a priori, especially seismic, data on the crustal structure the results of gravity inversion are encouraging. Essentially marine seismics (or even only Airy isostasy) and a "fixed datum" for Moho depth of Anatolia leads to a plausible model of the crustal structure of the eastern parts of the Black Sea and Anatolia and the transition with $<10\%$ uncertainty. With a rather simple crust-mantle structure including, beside the Moho, only thick marine sediments and a continental upper/lower crust boundary a rather "plausible" model has been derived. Moho depth varies between about 25 and 55 km under 2 km water and >2 km high terrain, respectively, with an uncertainty of only few kilometers. A more than 100 km broad region (between $x=190$ and 320 km) where topography is $\geqslant 2$ km has the deepest Moho; precisely speaking, the deepest Moho is found where the BA has its minimum (near $x\approx 320$ km), not where topography is highest (average >2200 m at $x\approx 190$ km) where in the model the Moho is transitional from sea to land. This rests on the assumption of lateral density homogeneity in crust and mantle; undulating lower/upper crust density variations and lateral density inhomogeneity in the mantle will modify the Moho.

The asymmetry implied may be related to a deep upper-mantle body of elevated density; such a body is postulated for reasons of isostasy: it allows the Moho to be about 3 km overcompensating the coast range topography. A limited (2D) mass of 100 km dimensions between 200 and 300 km depth and a density contrast of $+10$ kg/m^3 balances the shallower mass deficit relative to the Black Sea without influencing the higher level density contrasts in the inversion.

The mantle body of slightly elevated density may be the relict of former mantle lithosphere subduction and seems to be detected by tomography. Such a possibly detached slab may sink or be in a state of slow warming. These processes, if presently active, may be accompanied by an active uplift of the crust toward isostatic equilibrium; the high topography of the eastern Pontides may be an indication of such processes, but direct observational evidence is missing. The alternative model of lateral or regional isostasy (without the mantle density anomaly) which supposedly is inherited from Eocen rifting, some 50 Ma ago, is also compatible with gravity (Spadini et al., 1996). In view of the intervening complex plate tectonic evolution of the region it seems difficult to envisage such a heritage to be preserved till today.

The shorter gravity wavelengths (<100 km) can be "explained" by boundary undulations between an upper and a lower crust within a 10 km range. Such models have the effect that the sediment density contrast increases a little. In the case of a density increase with depth the undulations have the opposite sign to the case of a limited density decrease ($\sim 180°$ phase shift of undulations). Though lateral crustal density variations are most likely to exist there is no additional information about them.

The absolute values computed must be taken "with a grain of salt" as they systematically depend on a priori assumptions which have large uncertainties. Systematic errors should, however, not be worse than 10%. If the average mantle-crust density contrast is greater (smaller) than assumed, the average Moho depths "shrink" ("expand") accordingly. Thus, the Moho depth under eastern Anatolia should be between 50 and 60 km, under the eastern Black Sea (if the sediment thickness is correct) between 23 and 27 km. The models seem relatively reliable within the error limits and compatible with general geophysical knowledge and available seismic data. This is an example for the potential of gravity inversion with "reasonable", though sparse a priori information.

Acknowledgements

This study was done by cooperation between Trabzon and Mainz and largely depended on effective communication by e-mail. H.Ç and A.Ş are grateful for the opportunity to use the Mainz facilities and W.J. gives thanks for a pleasant visit to Trabzon Technical University. Carla Braitenberg did a thorough and thoughtful review which greatly helped improve the paper.

References

Andrusov, N.I., 1893. On the state of the Black Sea basin in Pliocene epoch. Geol. Paleont. St. Pet., Melang 1, 165–178 (in Russian).

Arkhangelsky, A.D., Strakhov, N.M., 1938. Geological structure and history of the development of the Black Sea. Izv. AN SSSR, 1938 (in Russian).

ARK Geophysics Ltd., no year. Gravity map of the Black Sea. Milton Keynes, England.

M.T.A., 1994. Genel Mudurlugu ve Harita Genel Mudurlugu, Ankara: Maden Tetkik ve Arama Genel Mudurlugu.

Barka, A.A., Kadinsky-Cade, K., 1988. Strike-slip fault geometry in Turkey and its influence on earthquake activity. Tectonics 7, 663–684.

Bektas, O., Yilmaz, C., Tasli, K., Akdag, K., Özgür, S., 1995. Cretaceous rifting of the eastern Pontide carbonate platform (NE Turkey): the formation of carbonate breccias and turbidites as evidence of a drowned platform. Giornale di Geologia, ser 3a (57/1–2), 233–244.

Belousov, V., Volvovsky, B.S., Arkhipov, I.V., Buryanova, V.B., Evsyukov, Y.D., Goncharov, V.P., Gordienko, V.V., Ismagilov, D.F., Kislov, G.K., Kogan, L.I., Kondyurin, A.V., Kozlov, V.N., Lebedev, L.I., Lokholatnikov, V.M., Malovitsky, Y.P., Moskalenko, V.N., Neprochnov, Y.P., Ostisky, B.K., Rusakov, O.M., Shimkus, K.M., Shlezinger, A.E., Sochelnikov, V.V., Sollogub, V.B., Soloviev, V.D., Starostenko, V.I., Starovoitov, A.F., Terekhov, A.A., Volvovsky, I.S., Zhigunov, A.S., Zolotarev, V.G., 1998. Structure and evolution of the earth's crust and upper mantle of the Black Sea. Boll. Geofis. Teor. Appl. 30, 109–198.

Bogolepov, A.K., Zhuravlev, V.A., Shipilov, E.V., Yunov, A., Yu, 1990. New data on the crustal structure under the Kara Sea, based on integrated geological and geophysical investigations. Transactions (Doklady) of the USSR Academy of Sciences: Earth Sci. Sect. 315, 112–116.

Cakir, Ö., Erduran, M., Cinar, H., Yilmaztürk, A., 2000. Foreward modelling receiver functions for crustal structure beneath station TBZ Trabzon, Turkey. Geophys. J. Int. 140, 341–356.

Cavsak, H., 1992. Dichtemodelle für den mitteleuropäischen Abschnitt der EGT aufgrund der gemeinsamen Inversion von Geoid, Schwere und refraktionsseismisch ermittelter Krustenstruktur. PhD thesis, Mainz.

Dobrynin, B.F., 1922. Geomorphology of the Crimea. Zemlevedeniye 1-2, 78–92. (in Russian).

Gardner, G.H.F., Gardner, L.W., Gregory, A.R., 1974. Formation velocity and density—the diagnostic basis for stratigraphic traps. Geophysics 39, 770–780.

Golmshtok, A.Y., Zonenshein, L.P., Terekhov, A.A., Shainurov, R.V., 1992. Age, thermal evolution and history of the Black Sea Basin based on heat flow and multichannel reflection data. Tectonophysics 210, 273–293.

Goncharov, V.P., Neprochnov, Y.P., Neprochnova, A.F., 1972. Bottom Relief and the Deep-sea Structure of the Black Sea Basin. Nauka, Moskow (in Russian).

Guterch, A., Grad, M., Materzok, R., Perchuk, E., 1986. Deep structure of the earth's crust in the contact zone of the Palaezoic and Precambrian platforms in Poland (Tornquist-Teisseyre Zone). Tectonophysics 128, 251–279.

Hearn, T., Ni, J., 1994. Pn velocities beneath continental collision zones: the Turkish-Iranian plateau. Geophys. J. Int 117, 273–283.

Jacoby, W.R., 1966. Schweremessungen auf Helgoland—Auswertung mit Ausgleichsverfahren. Z. Geophys. 32, 340–351.

Jacoby, W.R., 1973. Isostasie und Dichteverteilung in Kruste und oberem Mantel. Z. Geophys. 39, 79–96.

Kadinsky-Cade, K., Barazangi, M., Oliver, J., Isacks, B., 1981. Lateral variation in high-frequency seismic wave propagation at regional distances across the Turkish and Iranian plateaus. J. Geophys. Res. 86, 9377–9396.

Kropotkin, P.N., 1967. Mechanism of the Earth's crust movements. Geotektonika 57, 25–40. (in Russian).

Letouzey, J., Biju-Duval, B., Dorkel, A., Gonnard, R., Kristchev, K., Montadert, L., Sungurlu, O., 1977. The Black Sea: a marginal basin; geophysical and geological data. In: Biju-Duval, B., Montadert, L. (Eds.), International Symposium on the Structural History of the Mediterranean Basins. Editions Technip, Paris, pp. 363–376.

Makris, J., Stobbe, C., 1984. Physical properties and the state of the crust and upper mantle of the eastern Mediterranean Sea deduced from geophysical data. In: Morelli, C. (Ed.), Geological and Geodynamical Aspects on the Mediterranean Tectonics. Marine Geol. 55, 347–363.

Matthews, D.J., 1939. Tables of the velocity of sound in pure water and sea water for use in echo sound and sound ranging: (special issue). Hydrog. Dept. Ad. H.D. 383, Charts 62 a-b.

Milanovsky, E.E., 1963. Modern structure of the Caucasus and adjacent deep-sea basin as reflection of the earth's crust evolution studies in the Alpine geosynclinar Area. Vestnik MGU, Ser. Geol. 1, 48–58 (in Russian).

Milanovsky, E.E., 1965. The newest (late-orogenic) stage of geological development of the Caucasus (Neo-tectonics and the newest volcanism). Avtoref. Doktorskoy Dissert (in Russian).

Muratov, M.Y., 1955. History of tectonic development of deep Black Sea basin and its possible origin. Bull. Moip. Otd. Geol. 30, 27–50 (in Russian).

Nalivkin D.V., 1928. Paleogeography of the Central Asia in Paleozoic. In: Vsesojuzn. Syezda Geologoc; Tashkent (in Russian).

Obruchev, V.A., 1926. New tendencies in tectonics. Izv. Geol. Komissii 45, 117–140. (in Russian).

Okay, A.L., Sengör, A.M.C., Görür, N., 1994. Kinematic history of the opening of the Black Sea and its effect on the surrounding regions. Geology 22, 267–270.

Orovetskii, Yu, P., 1999. Mantle Plumes. Balkema, Rotterdam.

Ross, D.A., Degens, E.T., MacIlvaine, J.C., 1970. Black Sea: recent sedimentary history. Science 170, 163–165.

Ross, D.A., Uchupi, E., Bowin, C.O., 1974a. Shallow structure of the Black Sea-geology, chemistry, and biology. Am. Assoc. Petr. Geol. Mem. 20, 11–34.

Ross, D.A., Uchupi, E., Prada, K.E., MacIlvaine, J.C., 1974b. Bathymetry and microtopography of the Black Sea. In: Degens, E.T., Ross, D.A. (Eds.), The Black Sea-Geology, Chemistry, and Biology. Am. Assoc. Petr. Geol. Mem. 20, 1–10.

Sayil, N., Osmansahin, I., in press. Investigation of crust and upper-mantle structure at the Black Sea with group velocity data, Bull. Seis. Soc. Am.

Schindler, C., Pfister, M., 1997. Active Tectonics of northwestern Anatolia—The Marmara Ploy-Project. V/Dlf Hochschulverlag, Zürich. 570 pp.

Smilde, P.L., 1988. Verwendung und Bewertung von a-priori Information bei potentiell singulären Inversionsproblemen am Beispiel der gravimetrischen Bestimmung von Dichteverteilungen, Deutsche Geodätische Kommision, Reihe C, Heft Nr 490, München.

Spadini, G., Robinson, A., Cloetingh, S., 1996. Western versus eastern Black Sea tectonic evolution: pre-rift lithospheric controls on basin formation. Tectonophysics 266, 139–154.

Talwani, M., Worzel, J.L., Landisman, M., 1959a. Rapid gravity computation for two-dimensional bodies with applications to the Mendocino submarine fracture zone. J. Geophys. Res. 64, 49–59.

Wortel, M.J.R., Spakman, W., 2000. Subduction and slab detachment in the Mediterranean-Carpathian region. Science 290, 1910–1917.

Yilmaz, C., 1993. Evolution of the eastern Tauride carbonate platform during the Mesozoic: a sedimentological approach (middle-eastern Anatolia, Turkey). Giornale di Geologia, ser. 3a (55/2), 147–157.

Zonenshain, L.P., LePichon, X., 1986. Deep basins of the Black Sea and Caspian Sea as remnants of Mesozoic back-arc basins. Tectonophysics 123, 181–212.