

The Bouguer gravity map of the Mediterranean Sea (IBCM-G)

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Abstract. A Bouguer gravity map of the Mediterranean Sea and adjacent areas has been produced from a compilation of data gathered over a twenty-five-year period. An interpretation of the data is presented together with a description of the bathymetry and a gravity assessment of morphological features.

1. Introduction

Within the framework of the Intergovernmental Oceanographic Commission (IOC) and UNESCO International Bathymetric Chart of the Mediterranean (IBCM)/project, a set of 10 sheets at a scale 1:1 000 000 have been prepared for the Mediterranean Sea and adjacent areas, see Fig. 1 (Morelli and Val'chuk, 1988). The geophysical and geological overlay maps that accompany the bathymetry also include Bouguer gravity anomalies. The preparatory phase of the project and bathymetric surveys are illustrated in the IBCM explanatory leaflet. This paper presents a brief description of the gravity data and their processing and evaluation in Bouguer anomalies. Some density models, that satisfy the Bouguer field and are typical for the various tectonic provinces of the Mediterranean Sea, are also presented.

2. Data acquisition

2.1. Navigation

The Western and Central Mediterranean, from Gibraltar to approximately the 26°20'E meri-

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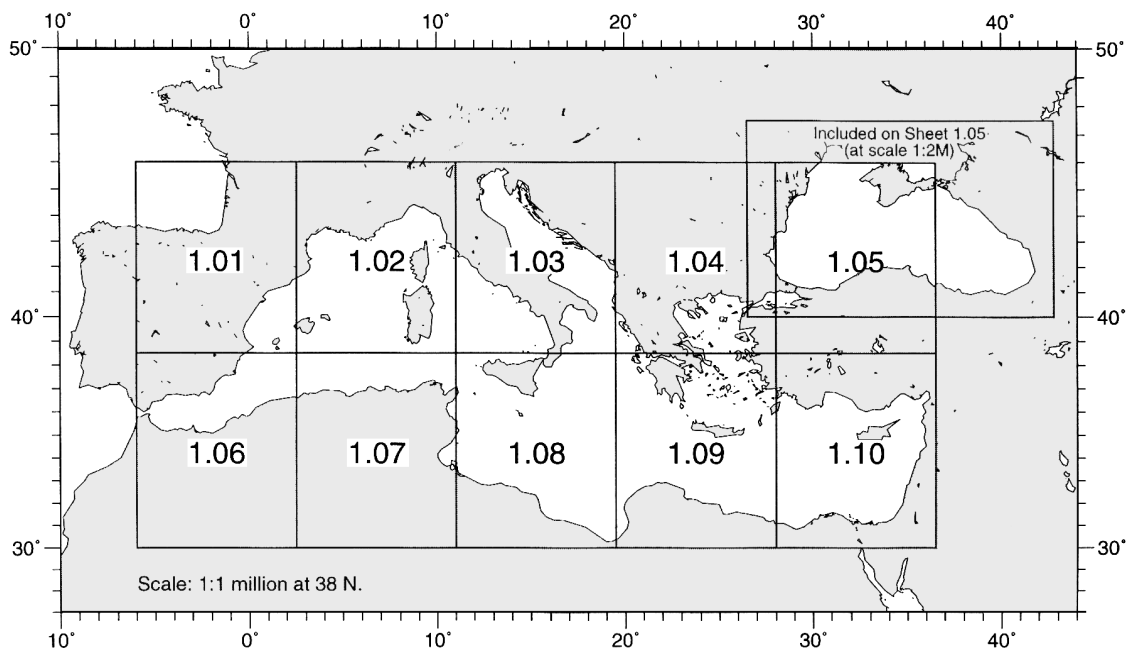


Fig. 1 - Plan of the sheets for the 1:1 000 000 International Bathymetric Map of the Mediterranean (IBCM).

dian, were surveyed between 1965 and 1972 by the Osservatorio Geofisico Sperimentale (OGS) at Trieste, using the *r/v Bannock*, and with financial support from the Consiglio Nazionale delle Ricerche, Rome. The survey extended to a total of 114 937 nautical miles (see Fig. 2).

Navigation was with the Loran C system. The geodetic elements of the transmitting stations were not available, but suitable tables of correspondence between Loran C and geographical coordinates were obtained, enabling a series of sheets with a plot of Loran C lines to be prepared. The scale chosen was 1:250 000 (Gantar et al., 1968) which provides a visual accuracy of 0.1' in geographical coordinates. In some marginal areas the signals received were highly biased, but periodic checks with radar fixes resolved most of the ambiguities. A constant speed between two successive Loran C fixes (~ 2.5 nautical miles) was maintained; corrections to the course were limited to less than 3° and spaced at intervals of at least one hour. In this way, preference was given to a continuous gravimetric record rather than to following an exact course.

The speed and course were computed according to the guidelines given by Boaga (1948, pages 190 and 223), on the international ellipsoid ($a = 6378388$ m, $e^2 = 0.0067226700$) and neglecting higher order terms in order to satisfy an accuracy of within 20 m for line lengths of the order of 100 km.

The data for the Levantine Sea are based on soundings from surveys carried out by the *r/v Shackleton* and *Discovery* in the period 1971-74, operated by the Dept. of Geodesy and Geophysics, University of Cambridge, both ships being satellite navigated (Woodside and Williams, 1977); and on soundings of less well navigated tracks made available to the International Hydrographic Bureau (IHB) by the Hydrographic Department, MOD, Taunton, UK (Wright et al., 1975).

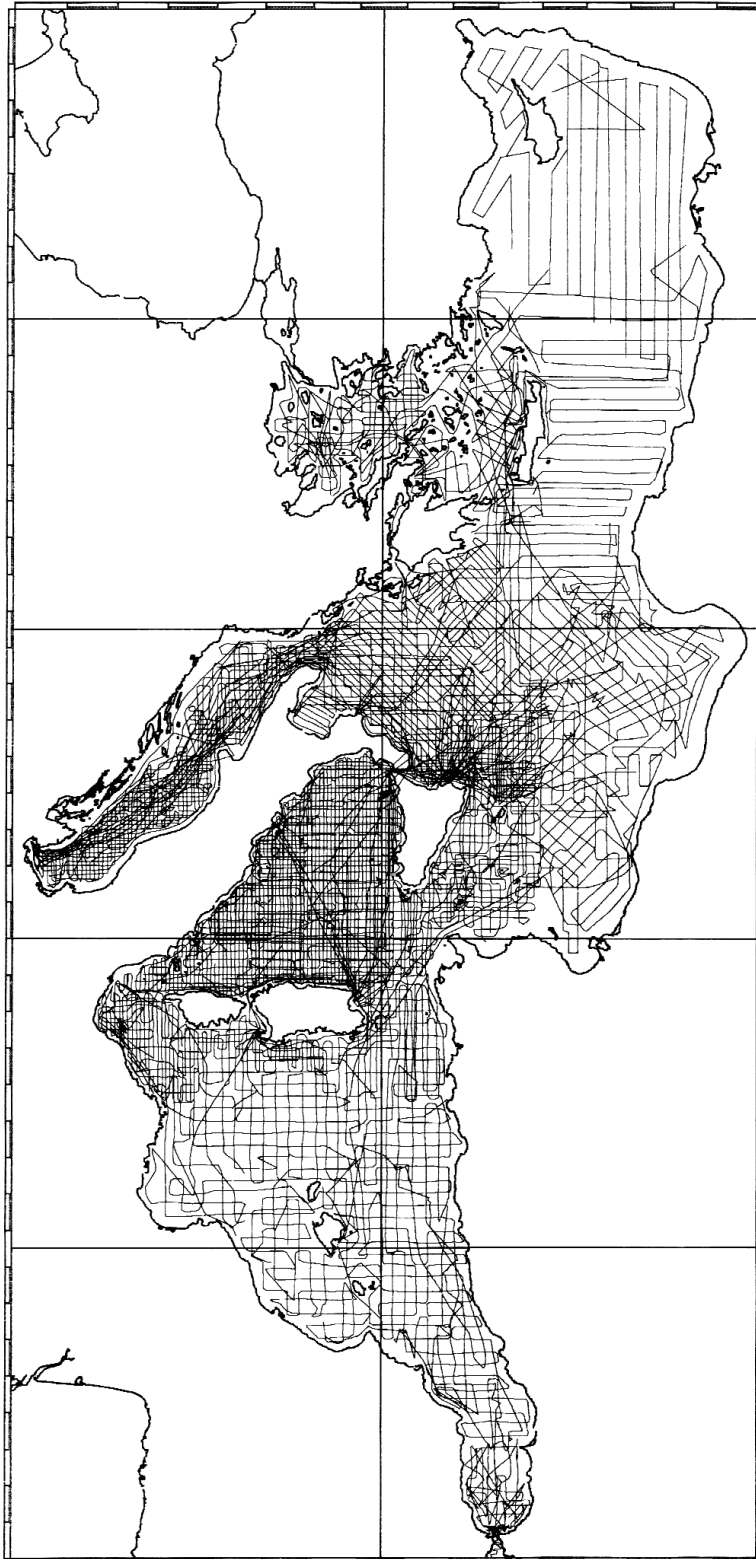


Fig. 2 - Tracks for the surface ship 1965-74 gravity surveys.

2.2. The gravity-meters

Gravity was measured using the Graf-Askania sea-surface Gss-2 gravimeters, n.13 and n. 11, mounted on Ausschütz gyro-stabilized platforms; sensitivity was about ~ 2 mGal/2.5 mm, and no reset was necessary. For an explanation of the theory of the Gss gravimeter, see Graf (1958), Worzel (1959), and Graf and Schulze (1961). The Askania-gravimeters were calibrated by means of readings taken in harbour before each cruise. These determinations also established the mean drift (~ 2 mGal/month, linear). Gravity values in the harbours were referred to Bad Harzburg (site 21510 A of the Bureau Gravimetrique International, BGI code, $g = 981180.40$ mGal).

The accuracy of observed gravity values, as determined from the track crossings, is ± 2 mGal. The accuracy of anomaly values is influenced by errors in positioning, interpolation, and in reduction. Interpolation errors depend upon the track spacings, variable from 3 km in the Central Mediterranean to 25 km or more in the Eastern Mediterranean. The tracks have been printed in muted color on the bathymetric sheets to allow users to evaluate the coverage. When considering reduction errors, the overall accuracy of Bouguer gravity anomalies on the surveyed profiles ranges from ± 3 mGal in the Central Mediterranean to ± 7 mGal in the peripheral areas. It has to be remembered that the measurements were performed in the years 1965-74, using surface ship gravity-meters of the 2nd series. With this degree of accuracy, correction for Earth tides can be neglected.

All values have been confirmed for the Eastern Mediterranean by comparison with satellite altimetry and other data (Arabelos and Tscherning, 1988).

3. Evaluation of the data as free air and Bouguer anomalies

The marine data set was integrated with the available land gravity data. The data were evaluated as free air values using the formula:

$$\Delta g' = g - \gamma_0 + \delta g_F + \delta g_E \quad (1)$$

where g is the observed gravity value; γ_0 is the theoretical value according to the Geodetic Reference System 1967 (GRS 68); δg_F is the free air reduction, δg_E is the Eötvös reduction obtained from

$$\delta g_E = 2wv \cos \varphi \sin \alpha + v^2/R \quad (2)$$

where w = velocity due to Earth's rotation; v = velocity of the ship; φ = geographic latitude; α = angle between geographic north and ship's course; R = Earth's radius.

The Bouguer values were obtained from

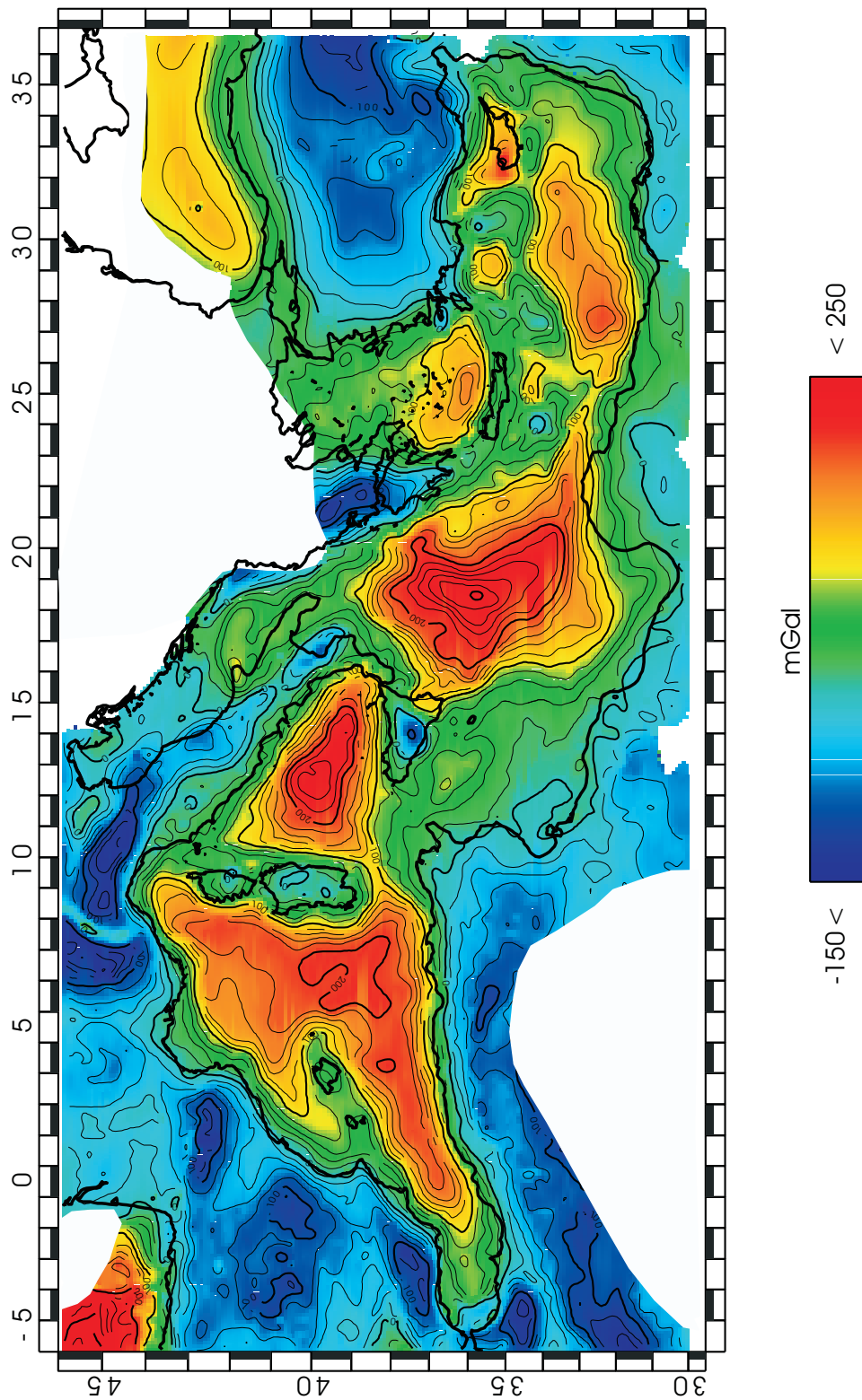


Fig. 3 - The Bouguer ($\Delta g''$) gravity anomalies in the Mediterranean area (equidistance: 10 mGal).

$$\Delta g'' = \Delta g' + \delta g_T + \delta g_B$$

where δg_T is the terrain reduction with a uniform density 2.67 g/cm^3 computed to Hayford Zone 0₂ on a spherical Earth (that is from 0 to 166.7 km); δg_B is the reduction of the Bouguer effect on a spherical Earth and to Hayford Zone 0₂. The formula used for this computation is that of Cassinis et al. (1937). The density used is also 2.67 g/cm^3 and the water density used is 1.04 g/cm^3 .

Land data on the Western Mediterranean countries have not been reduced for the effects of topography. Should the anomalies be used for geological modelling and quantitative interpretation, it is therefore necessary to be aware that the terrain effects in high mountains regions and steep coastal areas can easily reduce the accuracy by the order of 10 to 20 mGal. Instrumental drift and tidal effects have been considered and eliminated from the observed values.

4. Description of the Bouguer anomalies

The Bouguer gravity anomaly maps were published at the scale of 1:750 000 by Morelli et al. (1969) for the Adriatic sea; by Morelli (1970) for the Tyrrhenian sea, and by Morelli et al. (1975a, 1975b, 1975c) for the Strait of Sicily and the Ionian sea, the Aegean sea, the Eastern Mediterranean, and the Western Mediterranean. Most of the data for the IBCM and for the Bouguer overprint series at 1:1 000 000 scale have been obtained from this uniform series of maps, initially in analog form, and later digital.

The Bouguer anomaly IBCM-G sheets 1-3 and 6-8 (Fig. 1) were compiled in 1987 by the Bureau Gravimétrique International (BGI), Toulouse, France, from the original isoanoms by OGS. Sheets 4-5 and 9-10 were compiled by Makris at the University of Hamburg in 1988. The IBCM-G series was printed in 1989.

On land, three sources provided the bulk of the gravity values used in this present compilation. For most of the onshore areas of the Central and Western Mediterranean, gravity information was obtained from the BGI. The g -values are referred to the 1968 gravity datum and are listed together with their coordinates, j , l and h . The onshore data for the Eastern Mediterranean countries were prepared by the Institute of Geophysics, University of Hamburg (Makris, 1977) in gridded form from gravity values and maps available to the authors until 1985.

The sea gravity maps were analyzed in terms of their accuracy. Comparison between data collected in 1989 with the *r/v Meteor* using GPS-navigation and modern marine gravity meters, showed that the old surveys are good enough to permit 5 mGal-isolines to be drawn. The fact that the Western Mediterranean onshore data have no topographic reductions should be considered when interpreting.

Fig. 3 presents a simplified $\Delta g''$ map that covers the area of the 10 overlay sheets for the Mediterranean Sea and adjacent countries. In this map, which was generated 2 years after publishing the 1:1 000 000 scale map, the terrain effect for the Western Mediterranean countries was included. The shortcomings mentioned above for the first edition of the 1:1 000 000 Bouguer

maps are no longer relevant for the following discussion and models.

The values of gravity observed at the Earth's surface are strongly correlated with terrain morphology and hence with residual effects of tectonics, which are better preserved at sea than on land. In oceanic regions, gravity is also strongly correlated with crustal thickness and composition: it is generally lower than normal (negative anomalies) in sedimentary basins and higher than normal (positive anomalies) where the crust is thin and the oceanic denser mantle comes nearer to the surface. So, an initial assessment of the regional gravity anomalies offers a preliminary view of the structure of the Mediterranean area, which is characterized in almost all the deeper sea floor parts by strong to very strong positive Bouguer anomalies. From such an assessment, the Mediterranean Sea can be divided morphologically and gravimetrically into several sub-basins characterized by bathymetric depressions and gravity highs.

The eastern part is dominated by an elongated NE-SW trending depression, known as the Herodotus Abyssal Plain, with a Bouguer gravity maximum of approximately 220 mGal. The anomalies shoal to the east, and are separated from the eastern part of offshore Israel and Lebanon by less intense gravity anomalies ranging between 20 and 80 mGal. This decrease in the gravity anomaly level is caused mainly by an increase in sediment thickness of up to nearly 16 km at the Nile Cone and offshore Israel. The crust here is oceanic with a thickness of 5 to 6 km, an average velocity of 6.7 km/s and a density of 2.9 g/cm³. The Eratosthenes Seamount, on the other hand, is a continental fragment with a thickness of 24 km covered by relatively thin Mesozoic sediments, and has a prominent gravity expression (Makris and Wang, 1994). To the north, this broad and elongated gravity high is terminated by a number of gravity highs and lows that run more or less parallel to the continental areas of Turkey and Greece. Morphologically and geologically, this area falls within the Cyprus arc and the eastern part of the Hellenic arc. The gravity lows and highs coincide with bathymetric features. The Rhodes abyssal plain, for example, south of the island of Rhodes has a depth of approximately 4000 m and a gravity high of nearly 180 mGal.

In contrast to the Eratosthenes Seamount, the Anaximander Seamounts are marked by a gravity low of only 0 to 20 mGal. Unfortunately, very little is known about the deep structure in the area of these seamounts. It seems that changing thickness and density of the sediments alone cannot explain the rapid change in gravity level. Crustal thickening and change from continental to oceanic must also be involved. Whether the crustal elements observed are allochthonous parts of an accretional margin has still to be clarified by seismic sounding (Ben Abraham et al., 1987).

Cyprus, as part of the Cyprus arc, is exposed with a particularly strong positive gravity of nearly 220 mGal. Its crust, as Makris et al. (1983) showed by seismic soundings, is built from normal continental crust of approximately 30 to 32 km thickness. The area has been overthrust by Cretaceous ophiolites of considerable thickness (minimum 4 km) and exposed at the Troodos Massive. The mantle in all the areas described above is normal, as far as deep seismic soundings have shown; the Pn-velocity is of the order of 8.0 km/s (see also: Moskalenko, 1966; Woodside, 1977).

The Aegean Sea borders this region to the north-west and is located between two large complex continental areas; in the east by Anatolia, and in the west by the Hellenides. Both areas are marked by intense negative Bouguer values: -80 mGal in Anatolia and -120 mGal in western

Greece. In both areas the continental crust has a thickness of nearly 40 km and is covered partly by thick sediments. In western Greece, the sediments exceed 8 to 10 km thickness and are severely deformed by compression. The Aegean Sea between the two gravity lows is an uplifted continental crust thinned by stretching and covered by thin sediments, and with occasional small pull-apart basins containing thin sedimentary infill (Makris and Veis, 1977; Makris, 1978). The Cretan Sea is composed of stretched continental crust of only 18 to 20 km thickness at the 160 mGal gravity level, and includes 3000 m of Miocene and post-Miocene sediments.

The north Aegean crust thickens to near 26 km and in pull-apart depressions is overlain by up to 6000 m of sediments. In these transtensional pockets of the north Aegean Trough, the Bouguer gravity values range between +20 and +60 mGal. Negative $\Delta g''$ values are found only to the north of the coastal areas of the Rodopi mountains.

The Cretan arc has gravity values of between -10 to +30 mGal and separates the stretched Aegean domain from the Ionian and Libyan Seas. The deepest parts of the Mediterranean sea are located in the Ionian sea and Hellenic Trench. Similarly, the sediments accumulated in the accretionary prisms of the Mediterranean Ridge have slightly negative to positive $\Delta g''$ values, though the crust here includes relicts of oceanic basins. Such oceanic crust has been identified in the deep parts of the east and central Mediterranean from seismic soundings (Makris, 1981).

The largest positive $\Delta g''$ values, of some 300 mGal, coincide with the deepest part of the Ionian Sea next to the Mediterranean Ridge. The oceanic crust here is covered by only 4 to 5 km of sediments; the Pn -values are normal (8.0 km/s), with a density of approximately 3.35 g/cm³ (Makris et al., 1986).

In the Aegean sea, the observed relatively high heat flow values and the presence of a volcanic arc indicate that the upper mantle must be softened due to convective processes induced by the downgoing Ionian slab; Pn-values of 7.8 km/s and density of 3.2 g/cm³ have been reported by Lort (1971). To the west, the deep Ionian basin is bordered by the Calabrian arc its a fairly thick continental crust, overlain by thick sediments, and which accreted by crustal shortening and subduction. Sicily, Calabria and its extension towards the Apennines have negative Bouguer anomalies that range from -80 to -160 mGal in a south to north direction. The geometry and deformation of this negative gravity belt follows the compressional pattern between Italy and the Ligurian sea and oceanic Tyrrhenian Sea to the west. Active transcurrent faults displace the gravity lows from west to the east and can be traced eastwards below the Dinarides and the Albanian orogenic belt, which are also affected by crustal shortening and compression. Unfortunately, gravity values for the Balkan countries are unavailable for general use, and have not been incorporated in this edition of the maps.

The Calabrian Arc is limited to the west by the Tyrrhenian Sea, which for the most part has been oceanized in a transtensional regime. The Bouguer gravity values follow closely the geometry of this area and range between 180 mGal in the central part to 20 mGal on the bordering coasts of Italy, Sardinia and Corsica. They are associated with a thinned continental and oceanic crust ranging in thickness from 8 to 18 km. The Corso-Sardinia Block terminates this positive Bouguer field. The sediments thin in the central part of the Tyrrhenian Sea and thicken significantly in the bordering basins. The Corso-Sardinia Block consists of continental crust of average thickness (see Egger, 1992) of approximately 30 km and is covered by little or no sediment.

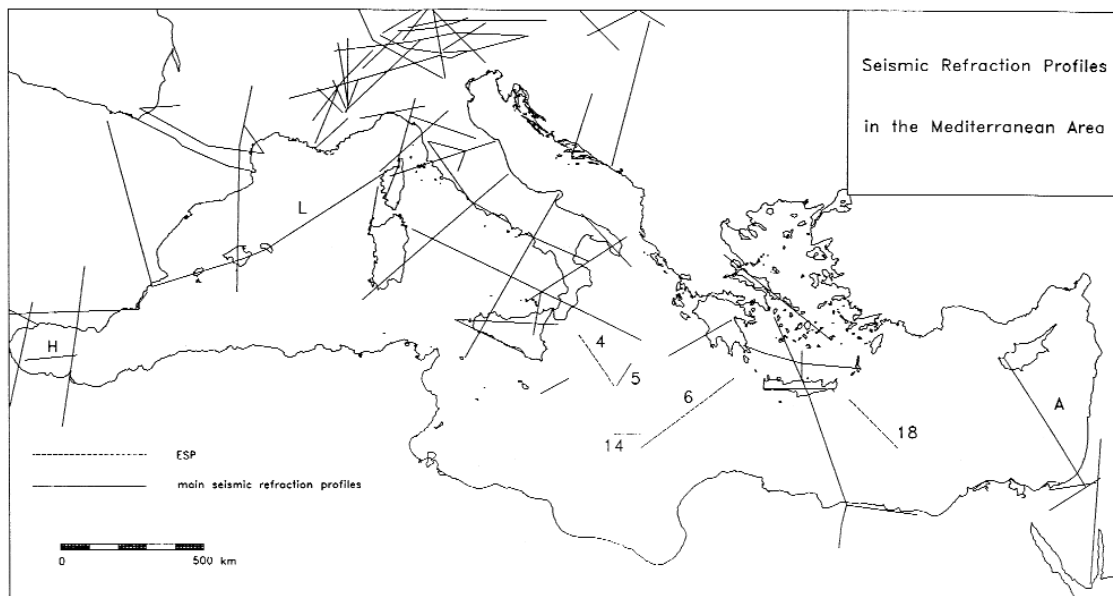


Fig. 4 - The main DSS profiles in the Mediterranean area and the ESP 1988 profiles (dotted).

The $\Delta g''$ gravity level is approximately -20 to -30 mGal. The Western Mediterranean basins west of the Corso-Sardinia Block are characterized by positive $\Delta g''$ values between 100 and 180 mGal. Most of the Ligurian Sea to the north, the Balearic Basin to the west and south-western part of this area are floored by oceanic crust, with typical $\Delta g''$ values above 140 mGal.

The Balearic continental splinter terminates the oceanic domain to the north at a gravity level of approximately 10 to 20 mGal, and as a 20 to 24 km thick continental crust with very little sediment (see Banda et al., 1980; Dañobeitia, 1990). The Spanish coast and the Balearic Islands are stretched continental crust of approximately 16 to 18 km thickness, covered by 6 km of sediments, thickening in near coastal basins. The gravity values follow the bathymetric features and change from approximately 100 mGal in the eastern part of the basin to $\Delta g''$ values of nearly 0 mGal at the Spanish coast.

The Alboran Sea is the western termination of the Mediterranean basin. It is floored by thin continental crust (see Hatzfeld and Frogneux, 1980; H in Fig. 4). The crust does not exceed 18 km in thickness, is covered by more than 6,000 m of sediments, and has $\Delta g''$ value ranging from -20 mGal at the Gibraltar arc to +40 mGal at its eastern end.

The continental areas that limit these basins to the south are parts of the African plate. They are characterized by strong negative $\Delta g''$ values associated with the Atlas mountains, and partly occupy areas floored by stretched continental crust covered by thick sediments, as seen in offshore Tunisia, in the Pelagian Sea and offshore Lybia. The crust thins in accordance with the $\Delta g''$ level from 30 km along the Atlas mountains to about 20 km in the Pelagian Sea. To the north, the Betics, Pyrenees and Alps mark zones of strong negative $\Delta g''$ values, which exceed -180 mGal in the central Alps, and are floored by thickened continental crust ranging between 38 km at the Betics, 38 to 40 km at the Pyrenees, and 50 km below the Alps (e.g., Blundell et al., 1992).

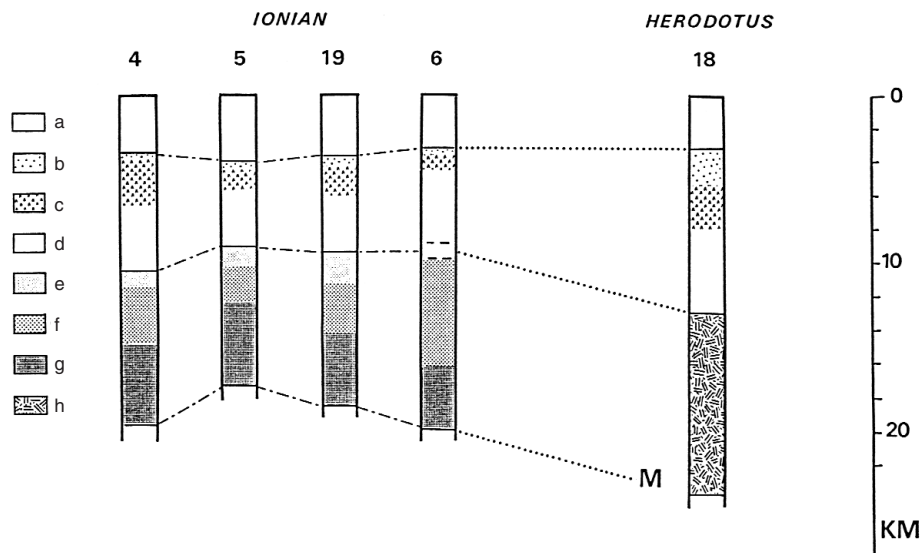


Fig. 5 - The ESP 1988 results (from de Voogd et al., 1992): a= water; b = recent sea-bottom sediments; c = evaporite; d= pre-Messinian sediments; e = oceanic layer 2a; f = oceanic layer 2; g = oceanic layer 3; h = oceanic or thinned continental crust; M = Moho.

The Variscan crust of Spain is at an approximate -80 mGal gravity level and is floored by normal continental crust of 30 to 32 km thickness. The Variscan crust of France is more variable in thickness, in gravity values that change in narrow belts between -20 and -70 mGal, and in thickness of crust and sedimentary cover. For crustal thickness maps see also Meissner et al. (1987).

In general, the gravity field of the Mediterranean Sea and the surrounding countries is controlled by their geological evolution. The strong negative $\Delta g''$ values are associated with the Alpine fold belts of northern Africa and Europe. Negative Bouguer anomalies are also associated with thick Variscan crust of the deformed parts of Mesozoic Europe. The deeper basins of the Mediterranean Sea, which are floored by oceanic crust, young or old, and covered by various thicknesses of sediments, have typical $\Delta g''$ values ranging between 100 and 300 mGal. The complexity of the gravity field on the expression of the complex deformation controlled by the transcurrent movement of Turkey to the west, and by the transcurrent movement of Africa and Iberia to the east and north-east. The eastern deformation belt is associated with the opening of the Red Sea and the relative movement of Arabia from Africa, whereas the deformation of Alpine Western Europe and Africa is influenced by the opening of the Atlantic.

The Black Sea to the north of Turkey has typical $\Delta g''$ values ranging between 20 and 100 mGal. It is floored partly by stretched continental and partly by oceanic crust, overlain by thick sediments. This area is also strongly affected by the presently active Alpine deformation, and is located between the Carpathian deformation and the Anatolian Wrench Fault to the south.

From correlation with bathymetric topography, the gravity anomaly field, shows that the broad geological elements are in isostatic balance, except for the zones controlled by strong compression, e.g., the external zones of the Hellenides (Makris, 1977), the southern Alps, the Apennines and parts of the Betics and Pyrenees.

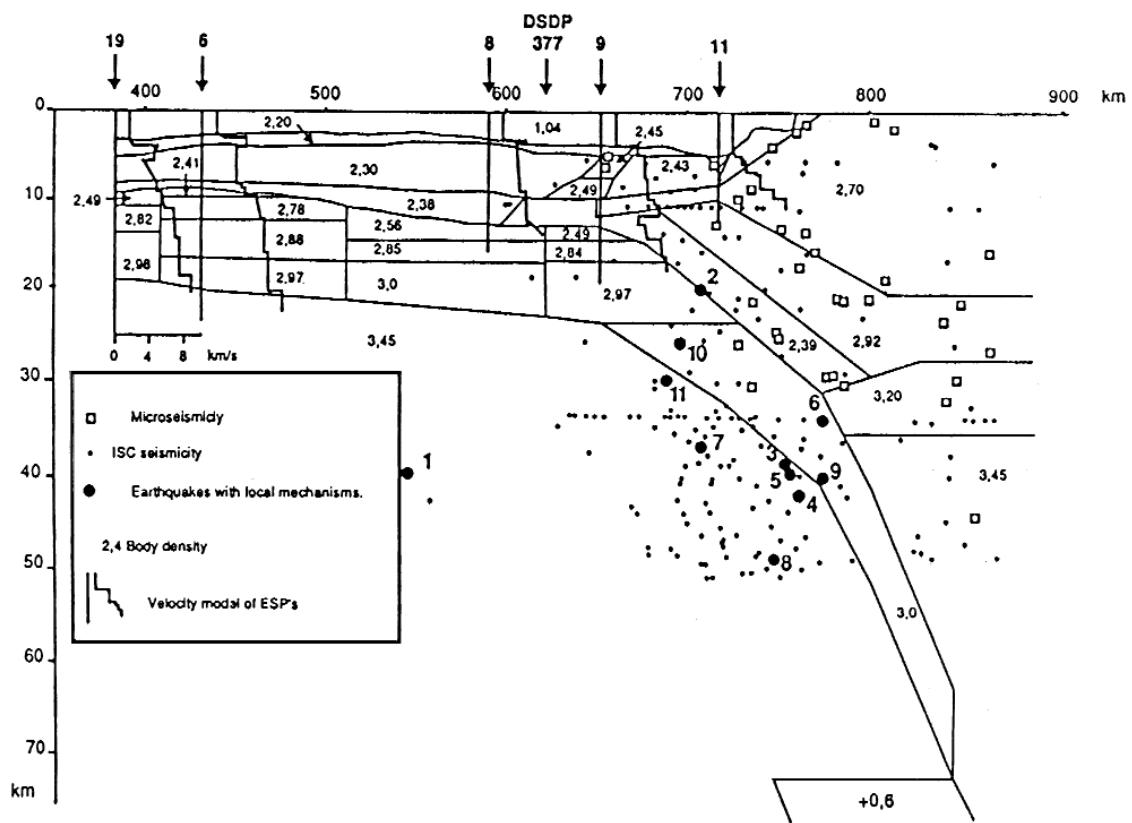


Fig. 6 - A gravity model for a profile NE-SW from the Southern Peloponnese through the Western Mediterranean Ridge (from Truffert et al., 1993).

5. Interpretation of the anomalies and density models

It is well known that inversion of gravity data requires constraints. These are normally imposed for the greater depths by seismic refraction (Deep Seismic Sounding = DSS; Fig. 4) or reflection data (Expanded Seismic Profiles = ESP). The nature of the crust in the deeps of the Eastern Mediterranean (being either oceanic or stretched continental) had long been a matter of lengthy discussions, due to its thickness and low heat flow. The problem was resolved in 1988 by a two-ship refraction and oblique deep seismic survey over the Ionian, Sirte and Herodotus abyssal plains (De Voogd et al., 1992; Fig. 4). These three areas have a relatively thin crust (8 to 11 km) overlain by a thick sedimentary cover, up to 10 km in the Herodotus abyssal plain (Fig. 5). The Moho boundary ($P_n = 8.4 - 8.5$ km/s) and the main crustal units identified in the basins can be followed beneath the Calabrian prism to the west, and beneath the Mediterranean Ridge to the east. The Moho progressively deepens towards the Hellenic Arc, in agreement with the accretionary prism model.

The crustal structure is oceanic for both the Ionian and Sirte basins, where typical oceanic layers, designated 2 and 3 in Fig. 5, are recognized. The thin crust of the Herodotus basin may be interpreted either as oceanic or thinned continental (about 10 km thick), but the top of the

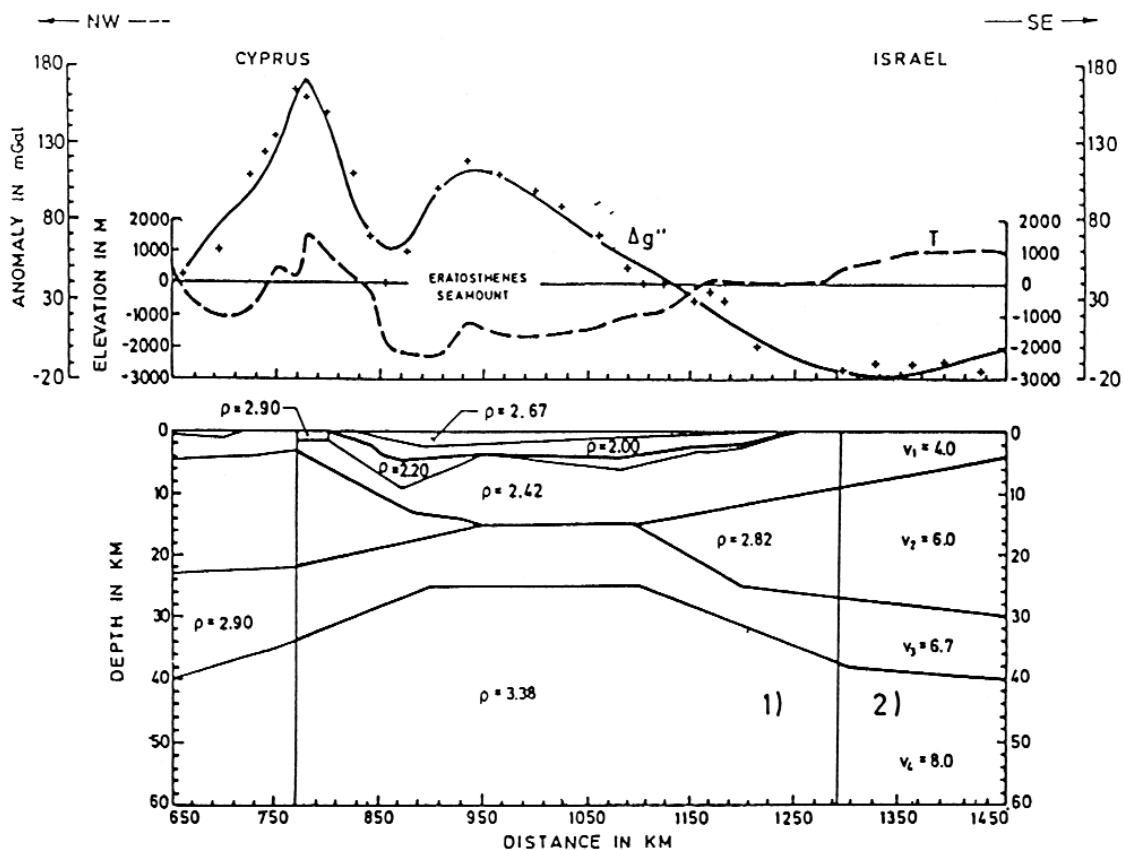


Fig. 7 - The Cyprus-Israel gravity profile constrained by DSS data (Makris et al., 1983).

crust in the Herodotus basin is very deep. Therefore, the Herodotus basin is probably significantly older than the Ionian basin, possibly Triassic as against early Cretaceous.

A crustal cross-section of the Western Mediterranean Ridge from the African margin to the Aegean Sea, constrained by the morphology of the ridge, new ESP results, the distribution of seismicity, and gravity modelling, was presented by Truffert et al. (1993). It was concluded that high velocity material occurs at shallow depths beneath the innermost unit of the ridge. Gravity modelling also indicates a significant increase in density beneath the inner unit (Fig. 6). It is suggested that the inner unit of the ridge which forms the backstop of the Hellenic Trench consists of continental material, possibly sheets of thrusts emplaced during the Oligocene to Miocene.

Crustal cross-sections from east to west typical of the evolution of the Mediterranean are now presented.

Fig. 7 shows a model through the Cyprus to Israel profile as computed by Makris et al. (1983). The $\Delta g''$ -model is seismically constrained, and VP values have been converted into densities using the Birch (1960, 1961) and Nafe-Drake (1963) empirical functions. The main features are a thickened continental crust under Turkey, bordered to the south by Cyprus, and an eastern Mediterranean Sea of oceanic origin with a thick sedimentary cover.

The transition between the thick continental block of Sicily and the western part of the Ionian

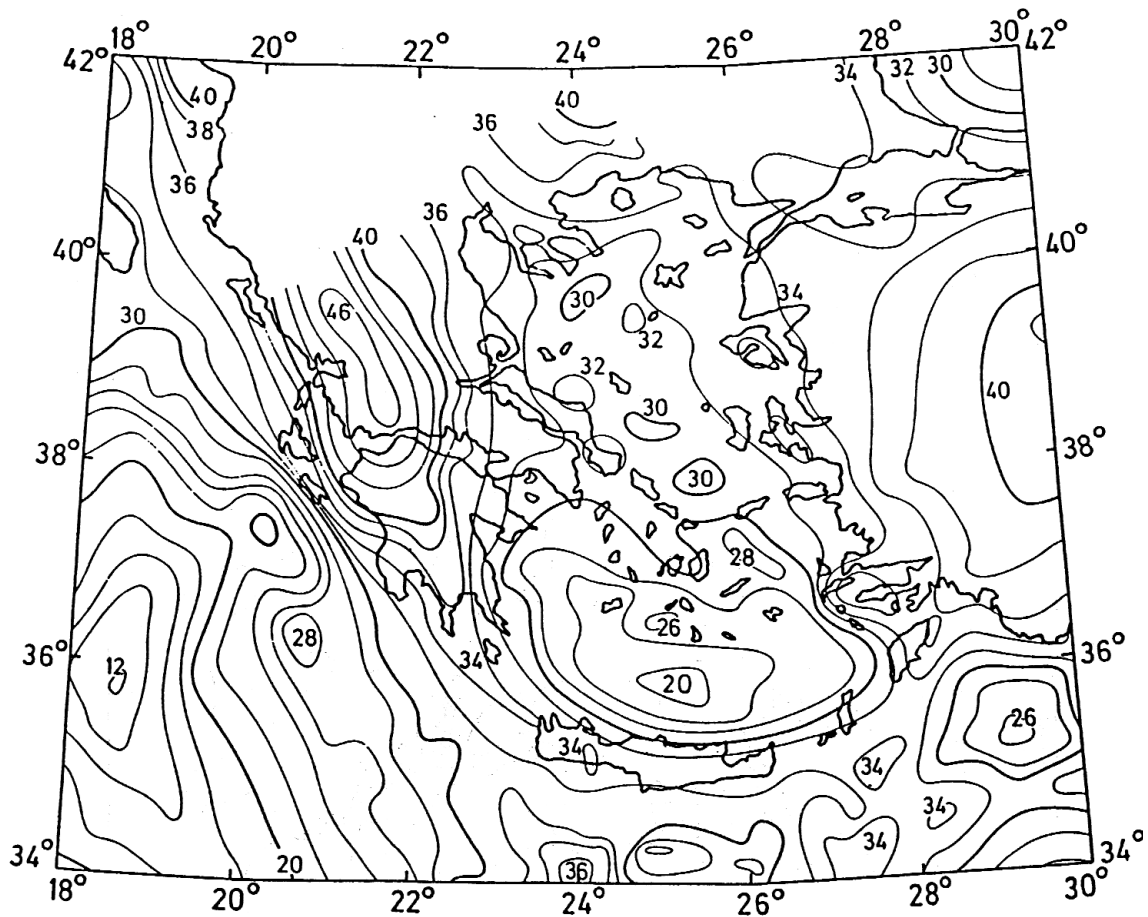


Fig. 8 - Moho map (equidepth: 2 km) deduced for the Eastern Mediterranean and adjacent regions from seismically constrained gravity data (Makris and Stobbe, 1984).

basin, still floored by stretched continental crust, is very abrupt (Makris et al., 1986). The central Ionian trough on the eastern part of the section is floored by oceanic crust, as discussed above. The style of deformation and distribution of densities shows that a simple model of stretching cannot satisfy the observed data: significant translational movement has deflected the tectonic units.

Fig. 8 presents a general view of the Moho depths in the Eastern Mediterranean computed from seismically constrained gravity data (Makris, 1981; Makris and Stobbe, 1984). In Fig. 9 the cross-section presented along the European Geotraverse (EGT) southern segment is reproduced as published by Klingel  et al. (1990). It can be seen how complex the deformation between the African and European plates is, and that the gravity field alone, without reliable constraints from seismic data, cannot produce a unique density distribution satisfying the observed gravity field.

Fig. 10 synthesizes our present knowledge of the Moho depths in the Central Mediterranean area, as deduced mainly from a detailed DSS net. These DSS data are very important for constraining and understanding the gravity anomalies. Some features of particular interest are:

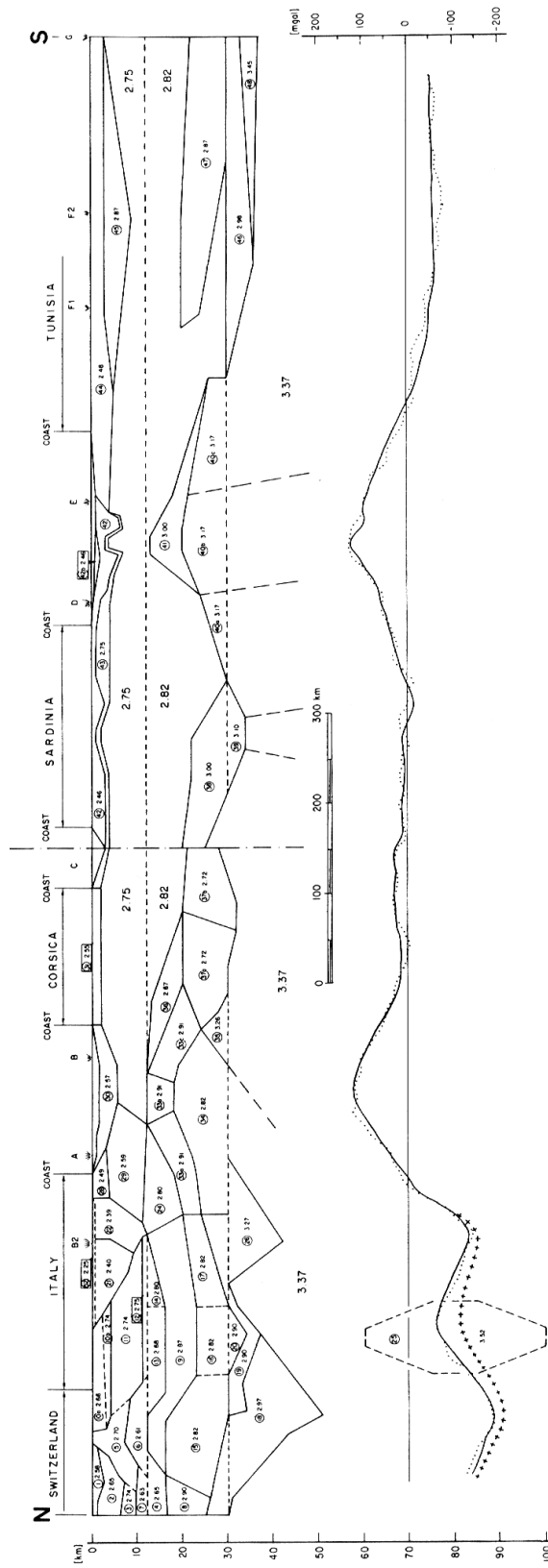


Fig. 9 - The gravity model of the EGT Southern Segment derived from seismic refraction data (Morelli and Nicolich, 1990). Solid lines represent seismic interfaces, while additional interfaces required by the gravity modelling are indicated by dashed lines. The separation between the northern and southern parts of the profile is marked by a vertical dash-dot-dash line. The lower part of the figure shows the observed Bouguer anomaly (solid line) and the computed one (dotted line). The crossed line shows the anomaly computed without the upper mantle disturbing body (from Klinge \acute{e} et al., 1990)

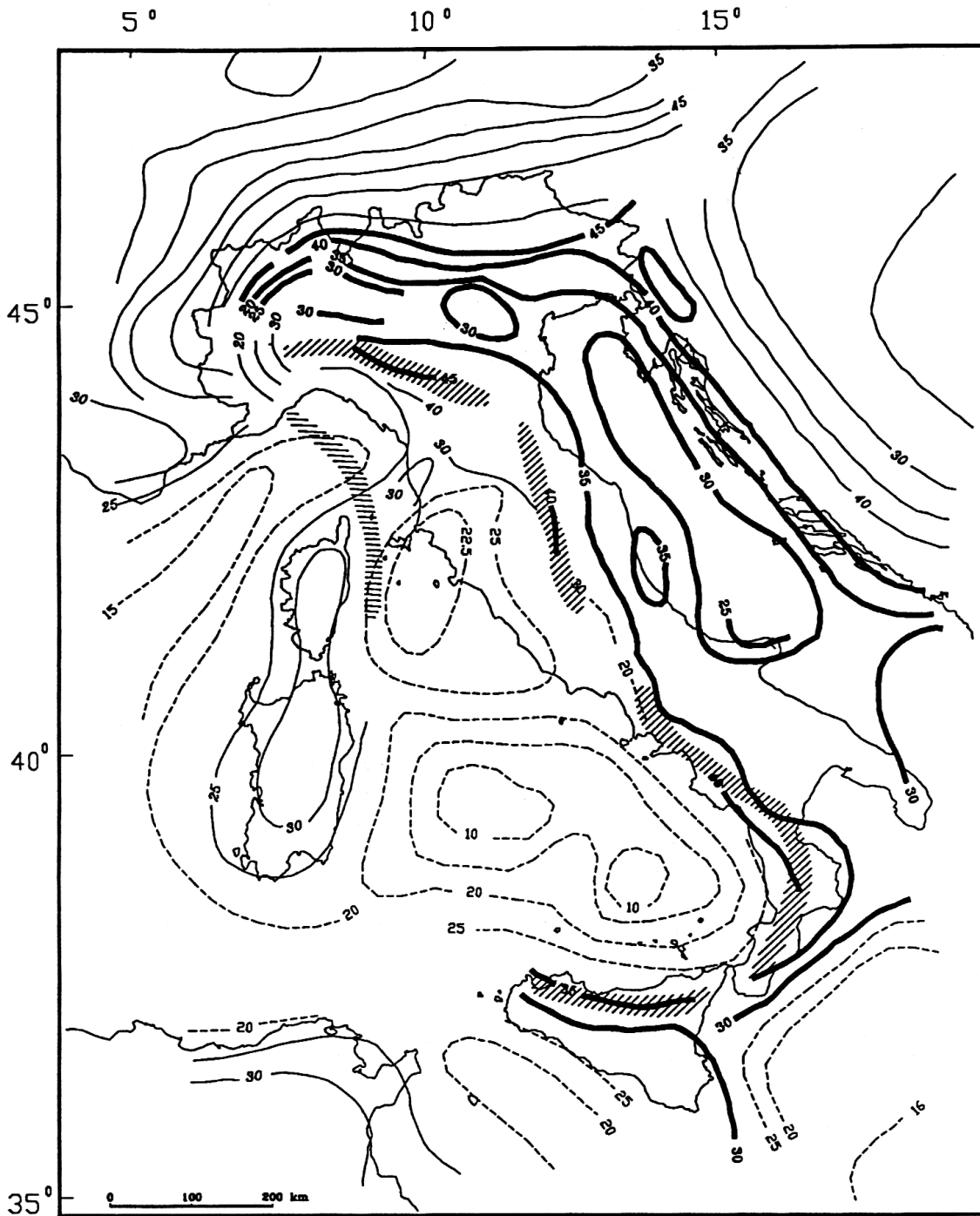


Fig. 10 - Moho isobaths (equidepth 5 km) deduced for the Central Mediterranean and adjacent regions from DSS data (from Nicolich and Dal Piaz, 1988, modified).

Moho isobaths: thick = Adriatic, thin = European, dashed = thinned (Tyrrhenic) or oceanic crust.

Hatchure: horizontal = maximal westwards extension of the Adriatic plate; inclined = contact and/or flexure (corresponding also to the maxima of the gravity gradient and of the seismicity) between the foreland normal Adriatic plate and its thinned part.

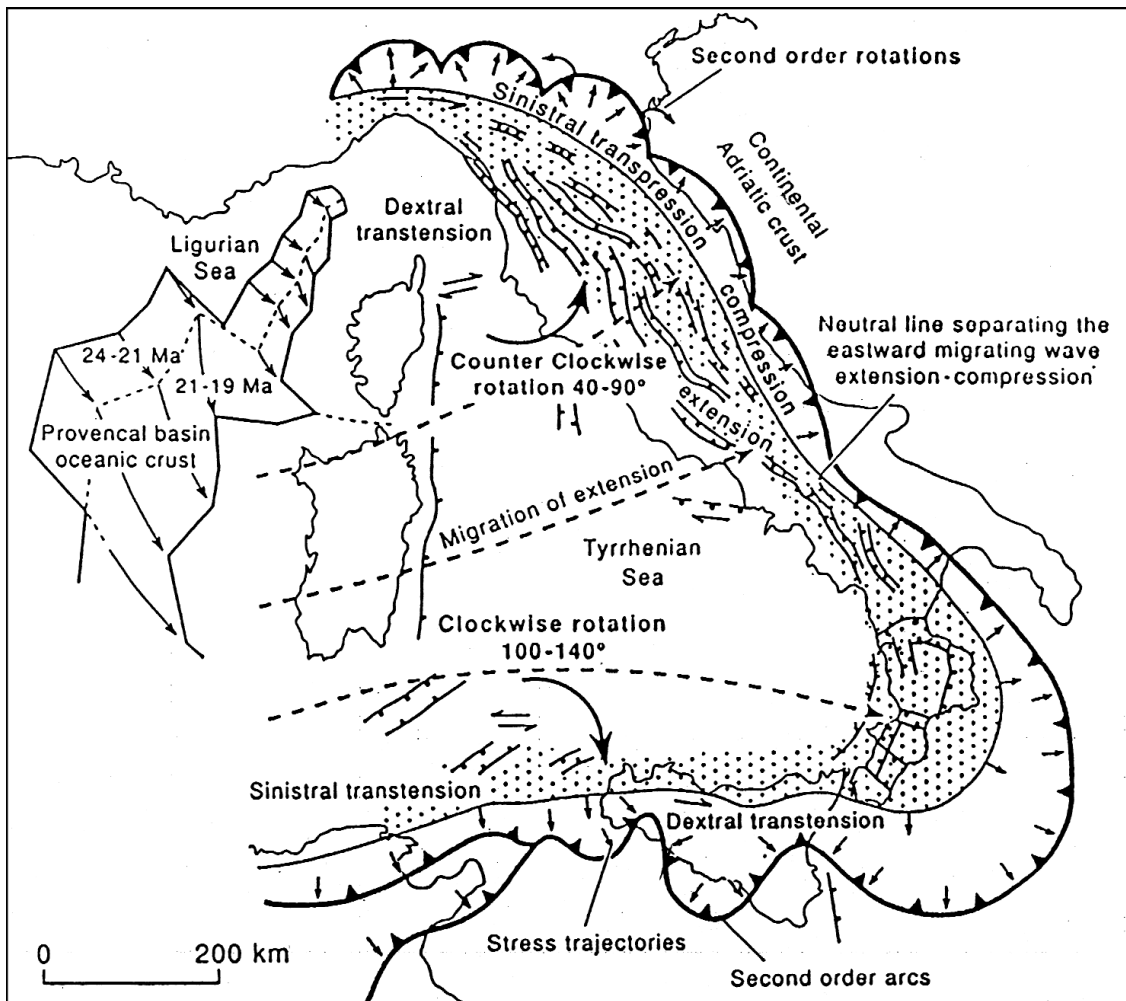


Fig. 11 - Kinematic model by Doglioni (1991) to show how the migrations and rotations of stress patterns have fashioned the present shape and state of the Tyrrhenian Sea tectonic system. The translation of Corsica and Sardinia from 24 to 21 Ma and their rotations between 21 and 19 Ma as the Provençal basin opened are shown after Burrus (1984).

- the crustal thickening and doubling in the collision zones, and the crustal thinning in the area of extension with delimitation of the Tyrrhenian oceanic windows;
- the bending of the continental ramps, which is mainly a consequence of the additional weight of the overlapping hinterlands (as in the Alps), both for the European plate (foreland) and for the Adriatic plate (hinterland), and on the western side of the Adriatic plate acting there as foreland, against the centrifugal thrusting from the Tyrrhenian area (hinterland);
- the Po-plain gravity maximum which can be attributed principally to the mantle dome under Milan.

The most probable origin of the centrifugal tectonic forces from the Tyrrhenian area are two ascending-descending mantle bodies (Locardi and Nicolich, 1988). At their borders, a revision of the seismicity in the area revealed two new important facts (Giardini and Velona, 1991), which

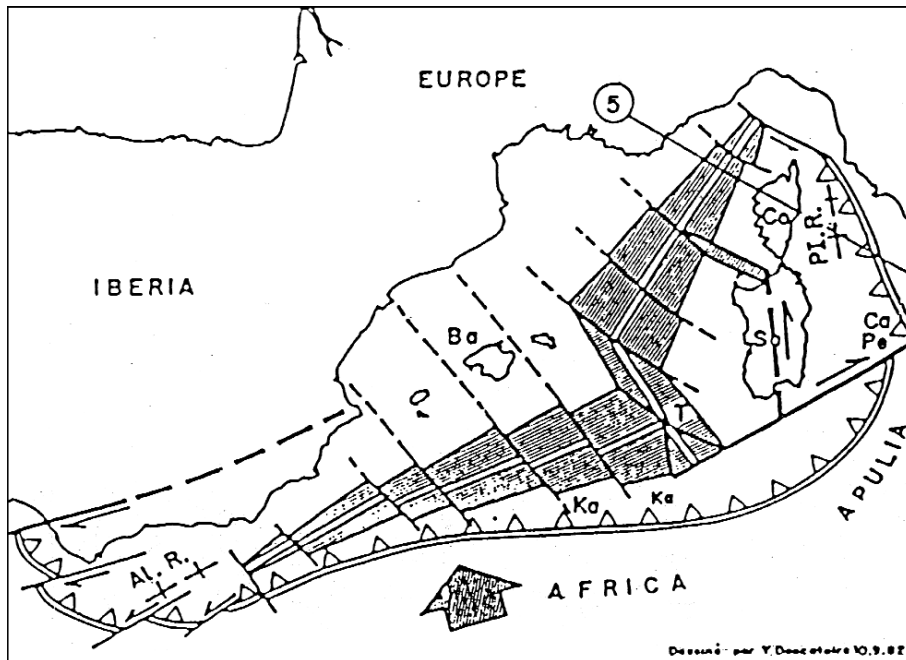


Fig. 12 - A kinematic sketch showing the Western Mediterranean at 18 Ma when drifting ceased and collision occurred along the northern African ranges and the Apennines (Rehault et al., 1984).

Ba, Balearic islands; Co, Corsica; Sa, Sardinia; Ca, Calabria; Pe, Peloritani structure; Ka, Grande Khabylie; ka, Petite Khabylie; P.L.R., Pianosa Ridge; Al.R., Alboran Ridge.

can be summarized as follows. Firstly, the peculiar traits of the stress geometry and the seismicity distribution indicate that the Tyrrhenian deep structure does not conform to the slab subduction model. Secondly, the travel time residuals reveal the clear presence of the slab below Sicily and the southern Apennines; however this pattern reverses in central Italy north of the Irpinia region, where raypaths do not 'see' the slab descending at depth.

The relocation by Selvaggi and Amato (1992) of 4700 Italian earthquakes 1983-1991, recorded by the new telemetered national network of the Istituto Nazionale di Geofisica, revealed 40 subcrustal quakes down to ~90 km in the Northern Apennines area, whose distribution is also in correspondence with the Tuscan mantellic body. The time sequence of the Corsica-Sardinia to Apennines drifting, in divergent directions, is very recent, see Fig. 11. The actual drift of the compressional fronts and the foredeep migration is 1.0-1.5 cm/y north-eastwards in the Northern Apennines; it is more than 5 cm/y southeastwards in the Southern Apennines (Scandone et al., 1990).

In the Western Mediterranean, the $\Delta g''$ and DSS data indicate a thin crust, which in the bathyal plain ($h \geq 2500\text{m}$) has been confirmed as oceanic. Indeed, Hirn et al. (1977) found in the Provençal Basin (along the profile L of Fig. 4) a Moho depth of 11 km and an uppermost mantle velocity of 7.7 km/s. But de Voogd et al. (1991) in a preliminary interpretation of the French part of the ECORS-CROP profile from the Gulf of Lions to Sardinia conclude that 'typical' oceanic crust cannot be readily recognised along the transect, though the oceanic nature of the cen-

tral basin is not questioned. The topographic characteristics of a spreading ridge are not observed, though features of uncertain nature are imaged within the basement at the northwestern end of the CROP line.

$\Delta g''$ and DSS data from the Alboran Sea to the Betics and the Variscan Meseta, indicate a deformation and almost crustal doubling of the continental material that has been compressed in the Betic chains. This intensive deformation influences not only the crustal geometry; the deep seismicity observed below the southern border of Iberia is also considered an expression of this deep-seated deformation and eventually subduction of early Alboran oceanic lithosphere, that has now been completely consumed.

6. Conclusions

The extensive work of observing, processing and presenting the gravity anomaly data has been presented here to open them to the scientific community. The description of the anomalies, their strong variations and characteristics have been illustrated with several examples. Integrating these data with other available geophysical and geological data, it is possible to summarise the very complex tectonic evolution of the Mediterranean area within the general framework of the oblique collision between Africa and Europe as follows:

- 1) the Eastern Mediterranean is a remnant of the ancient Tethys and, as a result of the African plate having broken against the European plate, is an area of active continental-continental collision in its easternmost (Levantine) part, and subduction beneath the Cretan, Ionian and Calabrian arcs;
- 2) the Pelagian and the Adriatic carbonate platforms east of the Apennine-Tyrrhenian arc belong to the African plate;
- 3) the rest of the Central, and all the Western Mediterranean were formed within the Alpine belt after the Cretaceous continental-continental collision between Africa and Europe. Starting in the west during the Oligocene (Alboran ~35 Ma) and continuing until the present (south-eastern Tyrrhenian, <2Ma), the crust was thinned and rifted by the tensional stresses in warmer bodies of the upper mantle, caused by subduction of the downgoing plate.

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