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A new high resolution geoid for the Iberian Atlantic continental shelf area

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Abstract. The first results of the new high resolution gravimetric geoid in the Iberian Atlantic continental shelf area are presented. Sea gravity data properly validated and covering the area $35° \leq \varphi \leq 45°$, $344° \leq \lambda \leq 353°$ are compared with satellite altimetry derived gravity anomalies in order to evaluate a $3' \times 3'$ gravity anomaly grid. The effect of the topography/bathymetry is taken into account through the RTM method and according to a remove-restore technique. The EGM96 is used as a reference field. Geoid heights are computed in the $3' \times 3'$ grid using the efficient 1D FFT procedure. In order to assess the quality of our computation in land from the gridded gravimetric heights, corresponding heights are interpolated onto GPS stations located on the mainland and along the west-coastal part of the Iberian peninsula as well. The standard deviation of the differences between the gravimetric and the GPS heights was found equal to 22 cm with a mean value equal to 55 cm.

1. Introduction

The sea gravimetric geoid is the fundamental tool for different geodetic and oceanographic studies. More specifically, it can be used in an indirect way for current estimation, sea surface topography determination in combination with altimetric information and for the computation of other quantities related to the gravity field applying inverse procedures.

Several years ago the Department of Geodesy and Surveying, of the Aristotle University of Thessaloniki, Greece, and the Instituto de Astronomia y Geodesia, (UCM-CSIC), Madrid, Spain, in the frame of a joint project carried out different studies related to the sea gravity field in different areas of the Mediterranean Sea (see e.g. Arabelos et al., 1994; Arabelos et al., 1996).

In continuation to these studies, a new high resolution geoid has been computed using sea

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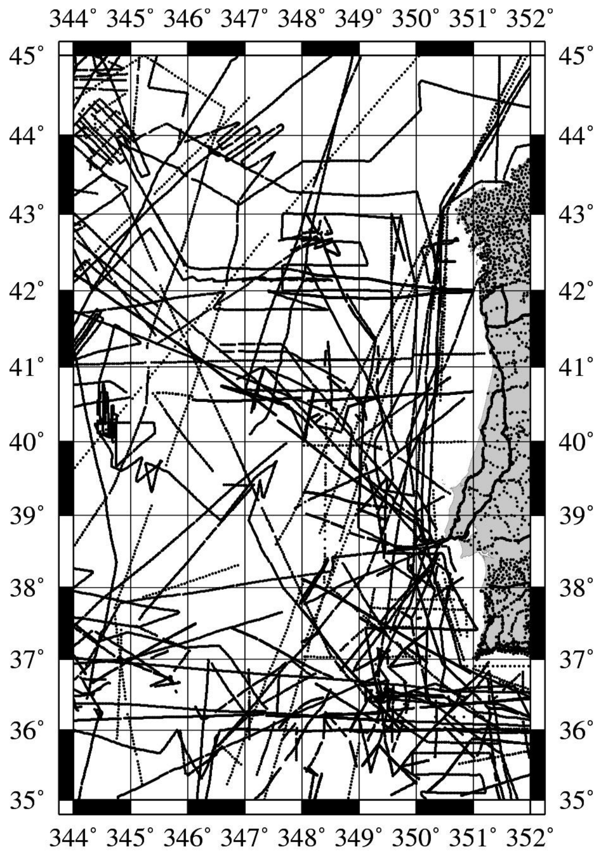

Fig. 1 - Distribution of Sea/Land gravity data.

gravity from various sea gravity surveys and using the recently developed EGM96 geopotential solution as reference.

Although the objective of this work is not that of presenting new methodological ideas, the new computed detailed gravimetric geoid may be used in a wide number of applications related to different branches of geosciences.

2. Data validation

The data sets used in this study were: (1) Geopotential models, (2) topographic/bathymetric data, (3) observed sea/land gravity data, and for comparison reasons only, gravity data derived from altimetry and (4) GPS/levelling heights.

2.1. Geopotential models

Two geopotential models were tested: The EGM96 (Lemoine et al., 1996) complete to degree

Table 1 - Statistics of the crossover differences of the observed gravity data (after the removal of outliers) before and after the crossover adjustemt. Unit is mGal.

	Mean	Std	Min	Max
Before the adjustment	0.60	12.50	-113.0	81.0
After the adjustment	0.52	4.25	-28.5	28.7

and order 360 and the ultra-high degree GPM98A (Wenzel, 1998) complete to degree and order 1800.

2.2. Topographic/bathymetric data

The ETOPO5U model was used after a shift correction. This correction was based on a validation with local data from the "Centro Geofisica Universidade de Lisboa" data bank. The shift - corrected ETOPO5U was compiled with new bathymetric data as described in (Catalao and Sevilla, 1997). The resulting model was used for the RTM reduction of the gravity anomalies.

2.3. Gravity data

The gravity data used in this study were extracted from different data banks. The sea gravity data were compiled from BGI, NGDC and DMA data banks. Most of the data are acquired from American, British and French Institutions in the period going from 1970 to 1990. The distribution of the data is shown in Fig. 1. The gravity anomalies were transferred to IGSN71/GRS80. Depending on the position, the distance between observations and the angle between each three consecutive observations, a number of 676 tracks were found. The gravity anomalies on each track were analyzed to detect internal outliers. This internal validation was performed by collocation. For each point, using 10 neighboring data points as input the intermediate gravity anomaly value was estimated and compared with the corresponding observation. When the difference between estimated and observed value was greater than 15 mGal, the observation was eliminated. Following this method, 387 observations were eliminated. The remaining 59 683 observations present a mean value equal to 14.65 mGal and a standard deviation equal to 48.68 mGal.

After a crossover analysis, 3992 crossover differences were detected. The statistics of these crossovers is shown in Table 1. Subsequently, a crossover bias adjustment was performed. The adjustment was carried out by fixing the longest track. Two different attempts were made by fixing (a) the longest track from the USA surveying and (b) the longest track from the UK surveying. The statistics of the gravity anomalies after the crossover adjustment showed a lower mean value when fixing the longest track from the UK surveying, while the standard deviation was the same for both cases. For this reason, the adjustment was finally based on the longest track from the UK surveying.

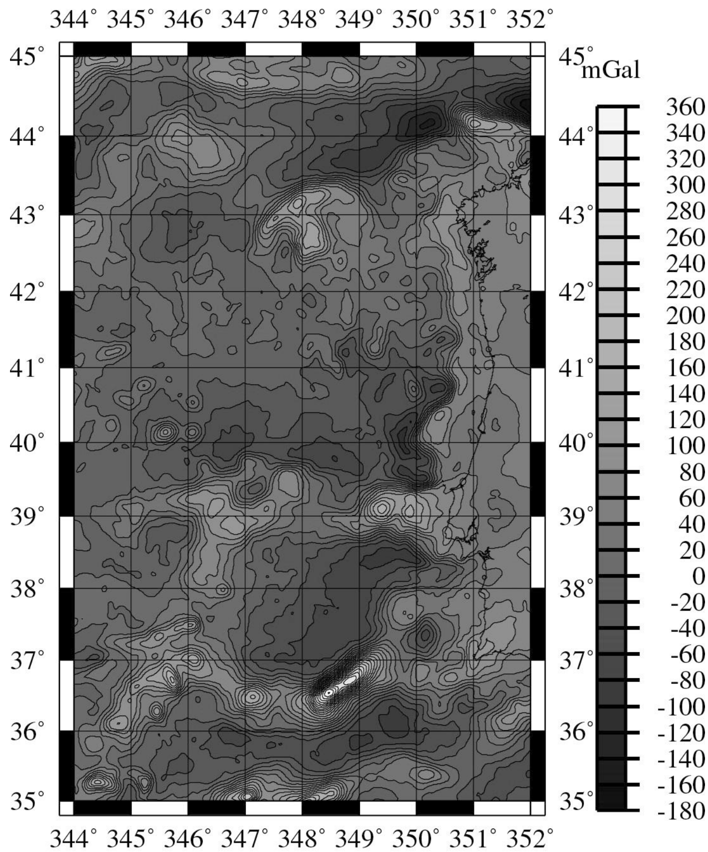

Fig. 2 - KMS-98 North-East Atlantic gravity anomalies.

Another problem was related to the treatment of the short tracks not connected to the main network. For the short tracks the same procedure as previously was applied. For each group of short tracks, one track, generally the longest, was fixed. These small groups of short tracks may introduce inhomogeneities in the final data set since they were processed separately. Finally, in order to detect unfitted data we decided to analyze each net individually, before accepting it.

In order to check the final data set for inhomogeneities, a comparison was made with gravity anomalies predicted from ERS-1 and GEOSAT altimeter data by Andersen and Knudsen (1998). The last data set consists of 2' × 2' gridded gravity anomalies and in the following will be referred to as KMS98 data set. The gravity anomaly field in our test area, based on KMS98 set, is shown in Fig. 2. The statistics of the differences between KMS98 and the validated final data set of the observed gravity anomalies is shown in Table 2. The corresponding differences are plotted in Fig. 3. From the statistics of Table 2 it is clear that, at least in terms of the mean value and the standard deviation, there are no systematic differences between the two data sets. Also, the pattern of Fig. 2 is very smooth with two exceptions: one in the lower part of the test area, close to the positive maximum anomaly of the gravity field, and another in the north-easten part of the test area, close to the north-western Iberian coast.

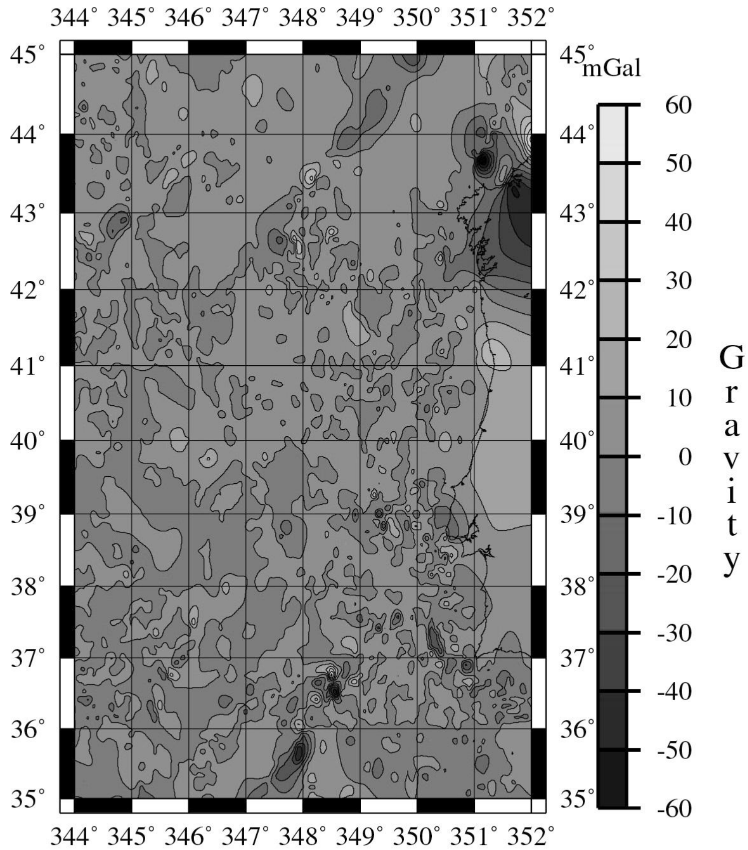

Fig. 3 - KMS98 minus observed gravity anomalies in the North-East Atlantic.

For the continental part of our test area, 7240 point gravity anomalies were available. The distribution of these gravity anomalies is also shown in Fig. 1. This data set was merged with the validated data set of sea gravity anomalies.

2.4. GPS derived heights

17 GPS derived heights were available along the west coast of Portugal. From this data set 9

Table 2 - Statistics of the differences between KMS98 gravity from altimetry and observed free-air gravity anomalies. Unit is mGal.

	Mean	Std	Min.	Max.
Observed (30,641 point values)	5.28	57.35	-168.80	345.85
KMS (2' × 2' grid, 76,941 values)	9.35	49.68	-184.08	360.11
Differences	0.19	6.53	-80.53	66.28

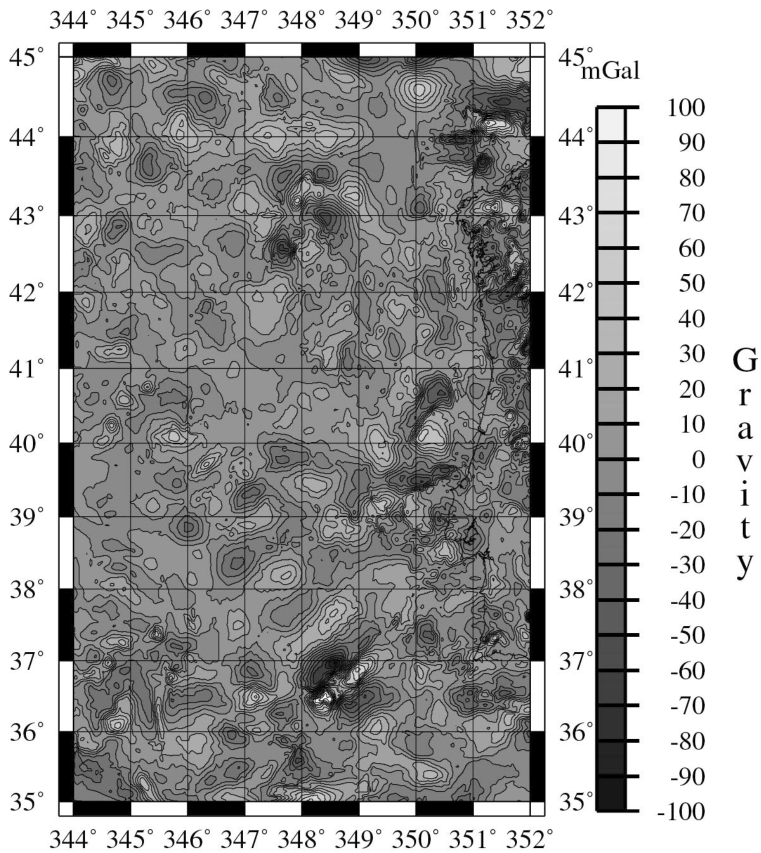

Fig. 4 - North-East Atlantic gravity anomalies minus EGM96 and RTM.

points were used for the comparison with the gravimetric derived heights. The remaining GPS stations lie outside our test area.

3. Computation of the geoid

For the geoid computation the usual remove-restore technique is used. The contribution of a geopotential model was subtracted from the gravity anomalies. Two geopotential models were examined: the EGM96 and the GPM98A. The statistics of the differences between observed gravity anomalies and those reduced to a geopotential model are given in Table 3 along with the statistics of the observed gravity anomalies. From the results in Table 3, it is obvious, that at least in terms of the standard deviation of the differences (observed - model), the GPM98A fits the gravity field in the test area very well. There is only one problem concerning the maximum value of the differences which in this case exceeds the maximum value of the observations. Such a large value in the differences was not found for EGM96 although the corresponding standard deviation is much larger (24.43 mGal). The EGM96 geopotential model was used in the end since this unexpectedly large value cannot be attributed to any obvious reasons.

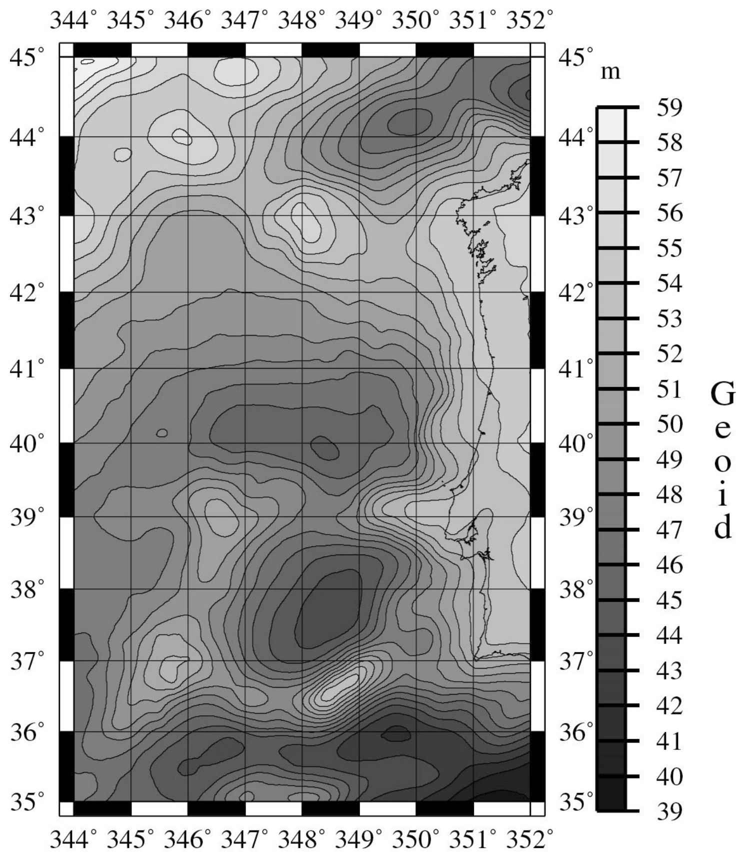

Fig. 5 - North-East Atlantic geoid.

The topographic/bathymetric data mentioned in section 2.2 were used for the RTM reduction of the EGM96 reduced gravity anomalies. After some numerical tests, better results, in terms of the mean value and the standard deviation for the RTM reduced gravity anomalies, were found using a 20' × 20' reference topography computed by averaging the 5' mean depths/heights and a value for the density equal to 2100 kg/m^3. The statistics of the RTM reduced gravity anomalies is shown in Table 3 too. The reduced gravity field is depicted in Fig. 4. From Table 3 we can see that the standard deviation of the reduced gravity anomalies is about 34% of that of the original ones. The strong gravimetric features of the gravity anomaly field are present in the observed

Table 3 - Statistics of the North-East Atlantic gravity anomalies (37 881 points). Unit is mGal.

	Mean	Std	Min.	Max.
Observed (37,881 point values)	13.433	55.403	-168.800	345.850
Observed-EGM 37,881 points	-0.2180	24.430	-122.890	169.510
Observed - GPM98A	1.0125	13.083	-88.150	354.887
Observed -EGM-RTM*	-0.484	18.882	-130.570	129.295

Table 4 - Statistics of the differences between geoid and GPS heights in the Western Iberian Peninsula. Unit is m.

	Mean	Std	Min.	Max.
geoid	54.27	0.82	53.17	55.23
GPS (9 points)	53.72	0.78	52.64	54.63
geoid - GPS (9 points)	0.55	0.22	0.12	0.86

gravity field (see Fig. 2) and in the reduced one as well (see Fig. 4).

To be able to compute the geoid using the FFT technique, from the trackwise distributed gravity values, a 3' × 3' grid was computed using kriging. The choice of the equidistance was based on the mean distance between points along track, as well as on the mean distance between neighbouring tracks. From the gridded gravity anomalies, the (residual) geoid was computed using the 1D FFT technique, which allows the evaluation of the true discrete Stokes' integral without approximation, parallel by parallel. The contribution of the EGM96 and the RTM effect were restored to the computed residual geoid heights. The computed geoid is shown in Fig. 5.

In order to assess the quality of our computation inland from the gridded gravimetric heights, corresponding heights were interpolated onto GPS stations located on the mainland and along the west-coastal part of the Iberian peninsula as well. The standard deviation of the differences between the gravimetric and the GPS heights was found to be 22 cm with a mean value of to 55 cm.

4. Conclusion

A high resolution absolute gravimetric geoid can be performed in land/sea areas by combining land/sea gravity data with topography/bathymetry data and high degree and order geopotential solutions. These regional geoid solutions can reach an accuracy of 10-20 cm.

For the future, the treatment of mean sea-level data from coastal stations is planned to obtain a better control over the land-sea interface. In considering the oceanic area, the gravimetric geoid heights will be compared with corresponding heights from the TOPEX/Poseidon altimetry mission. Additional comparisons with previous detailed geoid solutions available in the area will be carried out as well.

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