

Altimetry-Derived Marine Gravity Field Estimation Using Single- and Multi-Satellite Data

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Abstract. The possibility of improving the determination of the marine gravity field is investigated in an area offshore Newfoundland, Eastern Canada. Multi-satellite (ERS1, GEOSAT) geodetic mission (GM) altimetry data are used to improve the estimation of the gravity information inverted from altimetry. Newly estimated altimetry-derived local bathymetry models are implemented in the predictions aiming at providing as smooth residuals as possible before the gravity anomaly prediction takes place. The EGM96 geopotential model is used throughout this study to model the low-frequency part of the gravity field signal, while the altimetry data are corrected for the quasi-stationary sea surface topography (QSST) using the EGM96 dynamic ocean topography (DOT) model. Single- and multi-satellite altimetry-derived gravity anomaly fields are estimated and validated against shipborne gravity data and the KMS01 model. The estimation is carried out in the frequency domain using the efficient 2D planar FFT inverse Stokes convolution and employing discrete spectra for the kernel function. Special attention is paid to the modeling and removal of high-frequency oceanic phenomena contaminating GM altimetry through crossover adjustment and low-pass filtering. From the validation, it is shown that an altimetric gravity field accurate to about 3-5 mGal (1σ) can be estimated, while the use of multi-satellite data increases the resolution but does not manage to improve the final accuracy of the solutions.

Keywords. Altimetry, marine gravity field modeling, sea surface topography, sea surface variability.

1 Introduction

The estimation of a high-accuracy and high-resolution marine gravity field model is of high importance to most Earth sciences, since it provides useful information about the Earth's interior. During the last years numerous studies related to the use of altimetry data in marine gravity field modeling have been performed, all showing the great importance of implementing such datasets to improve the determination of the marine gravity field (Andritsanos et al. 2001; Hwang and Parsons 1998; Li and Sideris 1997; Sandwell and Smith 1997; Tziavos et

al. 1998; Vergos et al. 2001). All of these studies use Fast Fourier Transform (FFT)-based methods to derive the altimetric gravity field. Some of them invert directly the altimetric SSHs to derive gravity anomalies, while the rest derive first along-track deflections of the vertical from the SSHs and then use the inverse Vening-Meinesz formula to derive gravity anomalies (Hwang and Parsons 1998). Finally, some of them employ single-satellite data while others compute multi-satellite solutions, to improve the final resolution and accuracy of the estimated model (Andersen and Knudsen 1998).

The focus of this paper is to derive an optimal marine gravity field model, using single- and multi-satellite altimetry data in an area with high ocean dynamics and assess the achievable accuracy. Additionally, we want to investigate the influence of multi-satellite data on the final accuracy and resolution of the model. The final goal is to derive a unified approach/algorithm for altimetric marine gravity modeling implementing additional information such as bathymetry and QSST data in the processing procedure.

The effect of the ocean bathymetry in marine gravity field modeling can be taken into account through Digital Depth Models (DDMs) using the various topographic reduction methods such as the residual terrain modeling (RTM) reduction (Forsberg 1984). Their use aims mainly at providing smoother residual fields prior to gridding, interpolation and/or prediction. The effect of the QSST is important in processing altimetric data since the SSHs do not refer to the geoid but to the sea surface, thus their processing will determine a gravity field, which refers to the mean sea surface but not the geoid itself, i.e. a free-air gravity field. Thus, it is important to correct the altimetry SSHs for the QSST by simply removing its contribution.

The inversion of geoid heights to derive gravity anomalies is a differentiation that enhances the high frequencies, thus if the study is located in an area with high sea surface variability (SSV), then the resulting gravity field can be contaminated by noise. It is well known that the effect of the SSV appears in the densely spaced GM SSHs as high-frequency noise, which would be enhanced by the inverse Stokes if not removed/reduced. Thus, in the data processing, crossover adjustment and low-pass filtering are used to reduce the

SSV contaminating GM altimeter data and derive finally an optimal marine gravity field.

2 Altimetric Marine Gravity Field Modeling

Aiming at the determination of an accurate altimetry-derived marine gravity field model for the area under study, single- and multi-satellite solutions were determined. All solutions follow the same processing methodology, since their only difference lies in the combination of the multi-satellite data. This is performed by simple merging the irregular sub-satellite points into a combined dataset; then, the same processing concept is followed. Bathymetry as well as QSST data are implemented, while the well-known remove-compute-restore method is used.

Satellite altimetry data come in the form of SSHs that have to be corrected for the various geophysical effects and instrumental errors. After that step Corrected SSHs (CSSHs) are available from one or more satellites. Since for some (e.g. GEOSAT), observations refer not only to oceanic but land and shallow regions as well, a bathymetric mask has to be applied to remove the two latter. This is necessary since data over land and shallow regions contain in most cases errors due to the scattering of the radar pulse by dry land and the shallow ocean bottom and due to errors in the tide models close to the coastline. In the present study a depth equal to -50 m was selected to provide the oceanic observations. To the authors' opinion, the selection of this depth value is area dependant and its low value does not play a significant role, since if it is too small then the erroneous observations remaining in the data can be removed with a simple 3 rms test at a later step. The so-derived SSHs refer to oceanic regions only and have to be reduced from the sea surface to the geoid by removing the QSST. Most SST models come in terms of a spherical harmonics expansion of the DOT, so that the QSST can be computed for each SSH observation point as

$$\zeta_c(\phi, \lambda) = \left[\sum_{n=1}^{n_{max}} \sum_{m=0}^n (\bar{C}_{nm}^{*SST} \cos m\lambda + \bar{S}_{nm}^{SST} \sin m\lambda) \bar{P}_{nm}(\sin \phi) \right] (1)$$

where $\zeta_c(\phi, \lambda)$ is the contribution of the global DOT model coefficients, ϕ and λ denote latitude and longitude, where the model's contribution is determined, n_{max} is the maximum degree of expansion, $\bar{P}_{nm}(\sin \phi)$ are the fully normalized associated Legendre functions and \bar{C}_{nm}^{*SST} , \bar{S}_{nm}^{SST} are the fully normalized DOT spherical harmonic coefficients.

The so-derived SSHs refer now to the geoid and can be processed to give the final gravity field model using the remove-compute-restore method. Thus, the contribution of a geopotential model is removed to derive reduced SSHs. The so-referenced to a geopotential model SSHs may still contain some blunders, which should be removed, while the contribution of the bathymetry should also be taken into account. For the

blunder detection, a simple 3 rms test is used, which is sufficient to remove any erroneous observations still present in the data. One of the considerations at this point is that for a 3 rms test to be applied, all biases in the dataset should be removed, so that only random errors remain. This will be decided by examining the mean value of the reduced altimetry SSHs. If the mean value is low enough, e.g. below 10 cm, then we can proceed with the 3 rms test and then RTM-reduce the altimetry SSHs to derive the final residual SSHs. On the other hand, if the mean value is higher, then it means that some biases are still present in the data and should be removed prior to the 3 rms test. This is done by first RTM-reducing the SSHs, since when a good bathymetry model is used the resulting field is much smoother than the one prior to the reduction (Forsberg 1984).

The residual SSHs now available represent the medium wavelengths of the geoid height signal and can be safely regarded as residual geoid heights N_{res} . But, these measurements still contain the radial orbit error, due to the insufficient knowledge of the true satellite orbit, and the influence of time varying oceanic effects. Due to the improved orbit modeling of the latest altimetric datasets crossover adjustment may not be necessary for the reduction of the orbit errors, but may prove useful for the elimination of SSV-like effects (Knudsen 1992). This is true, since the height differences being adjusted at the crossover points contain neither the stationary geoid nor the QSST signals. Thus what mainly remains to be adjusted are, apart from orbit errors, unmodeled tidal phenomena and the SSH change due to SSV and other dynamic ocean effects. The present study has a local character, thus a regional crossover adjustment scheme with a bias and a tilt parameters (Rummel 1993) is sufficient.

After that step, the residual geoid heights (N_{res}) are gridded so as to predict gravity anomalies. To perform that, a weighted means with prediction power two type of gridding is used, taking into account the ten closest points for each grid node. This means that the inverse of the square distance of each point from the grid node is taken as its weight in the determination of the gridded N_{res} . To estimate the residual gravity anomalies (Δg_{res}) the contribution of the bathymetry (N^{RTM}) has to be restored to the N_{res} . Since the bathymetry refers to masses below the geoid, these have to be restored before the use of Stokes' formula for the determination of the boundary surface. The direct bathymetric effect should not be confused with the use of the terrain correction in the solution of boundary value problems using Helmert's condensation method (Forsberg 1984), where the effect of the topography is restored after the prediction of the residual gravity field. The Δg_{res} are estimated using the efficient 2D planar FFT inverse Stokes convolution and employing discrete spectra for the kernel function. Following Schwarz et al. (1997) and Tziavos (1995) this is given by a direct and an inverse FFT transform as:

$$\Delta g(x, y) = 2\pi\gamma \mathcal{F}_2^{-1} \left\{ \mathcal{F}_2 \{N(x, y)\} \omega \right\} \quad (2)$$

where \mathcal{F}_2 and \mathcal{F}_2^{-1} denote the direct and inverse 2-D FFT, $\omega = \sqrt{u^2 + v^2}$ is the radial wavenumber and γ the mean value of normal gravity. In all cases 100% zero-padding is appended around the N_{res} to avoid circular convolution effects. Due to the small cross-track spacing of GM altimetric data (3-4 km for GEOSAT), SSV, especially in open ocean areas, may not be completely eliminated by crossover adjustment, and will still be present in the gridded N_{res} . Thus, as much as possible of that high-frequency information, considered as noise in the signal, should be filtered out before deriving the final altimetric gravity anomalies. The filtering of the N_{res} is important, since the influence of the SSV will be enhanced by the inverse Stokes operator and can thus lead to gravity field estimates contaminated by noise. The reduction of the SSV is achieved by low-pass filtering the gridded N_{res} , using a collocation type of filter (Wiener filtering). This is performed in the frequency domain by multiplying the right-hand side of Eq. 2 with the filtering function $F(\omega)$:

$$\Delta G(\omega) = 2\pi\gamma\omega N(\omega)F(\omega) \quad (3)$$

The cut-off frequency ω_c (see Fig. 1) is empirically determined based on maximum noise reduction with minimum signal loss (signal to noise ratio). In that filtering operation different cut-off frequencies should be tested to select the one that provides the best N_{res} , based on the aforementioned criterion. The final step to determine the altimetric gravity field model is to restore the contribution of the geopotential model. The procedure described in this section is given schematically in Fig 1 (Vergos 2002).

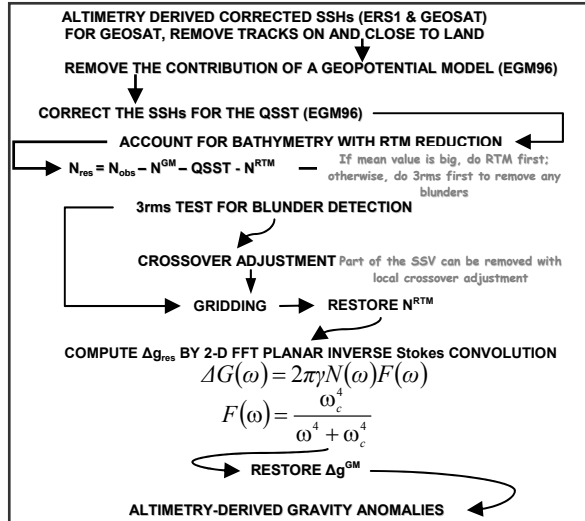


Fig. 1: Altimetric gravity field modeling.

3 Data Used and Gravity Model Validation

The area under study is located offshore Newfoundland, Eastern Canada bounded between $40^\circ \leq \phi \leq 50^\circ$ and $310^\circ \leq \lambda \leq 320^\circ$. ERS1 and GEOSAT GM satellite altimetry data from the latest releases of their GDR's have

been extracted from the databases of AVISO (1998) and NOAA (1997) respectively. The local depth models used to take into account the effect of the bathymetry were those developed by Vergos and Sideris (2002) and Vergos (2002) using satellite altimetry and depth soundings. Finally, the EGM96 geopotential model, complete to degree and order 360, and the EGM96 DOT, complete to degree and order 20, were used to provide the long wavelength geoid information and the QSST estimates respectively (Lemoine et al. 1998).

The validation of the estimated gravity models is performed through comparisons with 97474 shipborne gravity data provided by BGI and the Geodetic Survey Division of Natural Resources Canada (see Fig. 2) as well as with the 2'x2' KMS01 global altimetry-derived gravity field model (Andersen and Knudsen 1998). In all cases the differences were formed in the sense $N^i - N^j$ where i represents the estimated single- or multi-satellite altimetric gravity model and j denotes the shipborne or KMS01 data used for the validation

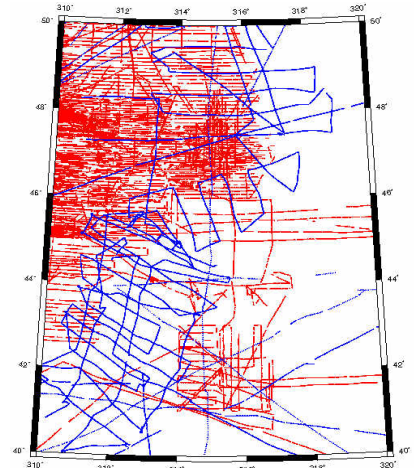


Fig. 2: Shipborne gravity data distribution used for the validation.

4 Development of Gravity Field Models

The input data consist of 42640 ERS1 GM and 76485 GEOSAT GM satellite altimetry observations. Local crossover adjustment of the satellite arcs is applied only for the case of GEOSAT, since the ERS1 data have been already adjusted by AVISO. Table 1 presents the statistics of the GEOSAT residual geoid heights (N_{res}), where it is shown that the field after the adjustment has a smaller σ by 3 cm, compared to the one prior to crossover, while the mean value is also smaller by 1 cm. The estimated bias and tilt parameters are equal to 2.1 cm and -0.9 cm for the ascending arcs and -2.5 cm and 0.3 cm for the descending ones. The small reduction of the mean value after the adjustment indicates that the new orbits of the altimetry data are determined very accurately so that crossover adjustment may not be necessary for the reduction of the radial orbit error but for the elimination of other effects.

After crossover adjustment the next step refers to the use of bathymetry to smooth the residual fields. Both

the GEOSAT and ERS1 residual fields have been RTM reduced using the local bathymetry developed by Vergos (2002). Table 2 presents the statistics of the RTM-reduced N_{res} where a 3 rms test for blunder removal has been applied as well. For GEOSAT, the blunder detection and removal was performed prior to the RTM-reduction since the mean value was equal to -0.08 cm only. For ERS1 the 3 rms test was performed after the RTM-reduction, since the mean value prior and after that was equal to 0.21 and 0.19 cm respectively.

Table 1. Statistics of GEOSAT N_{res} before and after the crossover adjustment. Unit: [m].

	max	min	mean	σ
before crossover	0.85	-0.84	0.13	± 0.23
after crossover	1.19	-1.00	0.12	± 0.20

Table 2. Statistics of GEOSAT and ERS1 N_{res} before and after the RTM-reduction. Unit: [m].

	max	min	mean	σ
before (GEOSAT)	1.19	-1.00	0.12	± 0.20
after (GEOSAT)	1.09	-0.98	0.10	± 0.18
before (ERS1)	1.11	-0.99	0.21	± 0.25
after (ERS1)	0.97	-0.95	0.19	± 0.22

For both satellites it is interesting to notice the improvement of both the mean and σ values after the RTM-reduction as well as that of the range of the N_{res} . These are reduced by about 2, 2 and 12 cm for GEOSAT and about 2, 3 and 18 cm for ERS1 respectively, showing that the resulting fields after the RTM reduction are indeed smoother. Another important point is that when the global JGP95E (Lemoine et al. 1998) bathymetry model was used, the residual fields became much rougher in both cases with the σ increasing to ± 1.60 m for GEOSAT and ± 1.65 m for ERS1 and the range to more than 7 m for both satellites. That is mainly attributed to the model's insufficient resolution (5') and errors (see the analysis in Vergos 2002; Vergos and Sideris 2002). Thus, the use of bathymetry in marine gravity field modeling should be performed with caution, since an inaccurate model can lead to a less accurate estimation of the final field.

The data were then gridded on a $3' \times 3'$ grid to form the N_{res} used for the prediction of residual gravity anomalies, while the combined dataset was created by simply merging the irregular single-satellite N_{res} and then by gridding it. The contribution of bathymetry was restored to all datasets and then the prediction took place. To choose the optimal cut-off frequency for the low-pass filter, different selections were tested for each of the three datasets. The final ω_c 's were set to 20 km for the ERS1, GEOSAT and combined datasets. Table 3 presents the statistics of the estimated Δg_{res} for the three solutions. To derive the final single- and multi-satellite models, the contribution of the EGM96 geopotential model was restored, resulting in the fields whose statistics are presented in Table 4. Figs. 3 and 4 depict the GEOSAT and combined gravity fields respectively. From these figures it can be seen that the combined solution provides a more detailed representation of the

gravity field in Newfoundland especially over the Milne ($44^\circ \leq \varphi \leq 46^\circ$ and $318^\circ \leq \lambda \leq 320^\circ$) and Newfoundland Seamounts ($\varphi = 44^\circ$ and $313^\circ \leq \lambda \leq 317^\circ$). Additionally, both solutions manage to depict very well the positive gravity anomalies over a seamount ($\varphi = 47.75^\circ$ and $\lambda = 318.5^\circ$) that was identified by the bathymetry from Vergos (2002) but not represented by JGP95E bathymetry and only roughly depicted by KMS01. Finally, the Flemish cap (the plateau feature in the north-central part of the area) is clearly distinguished.

Table 3. Statistics of predicted GEOSAT, ERS1 and combined Δg_{res} . Unit: [mGal].

	max	min	mean	σ
GEOSAT	54.9	-56.0	0.04	± 10.1
ERS1	70.7	-46.7	0.04	± 10.8
Combined	71.7	-58.0	0.04	± 10.5

Table 4. Statistics of the final GEOSAT, ERS1 and combined gravity fields. Unit: [mGal].

	max	min	mean	σ
GEOSAT	136.4	-61.6	14.1	± 23.1
ERS1	132.2	-66.3	14.1	± 23.2
Combined	134.8	-65.7	14.1	± 23.1

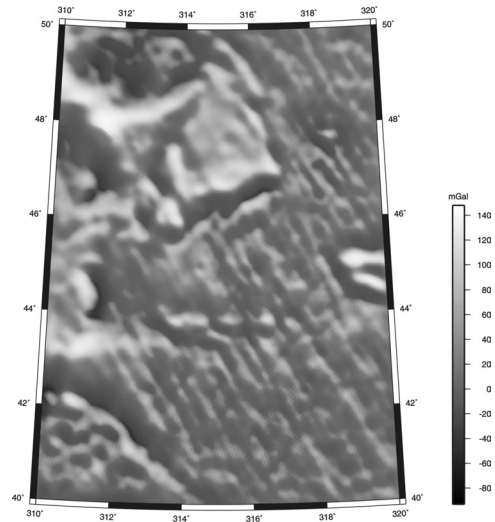


Fig. 3: GEOSAT single-satellite gravity field in Newfoundland.

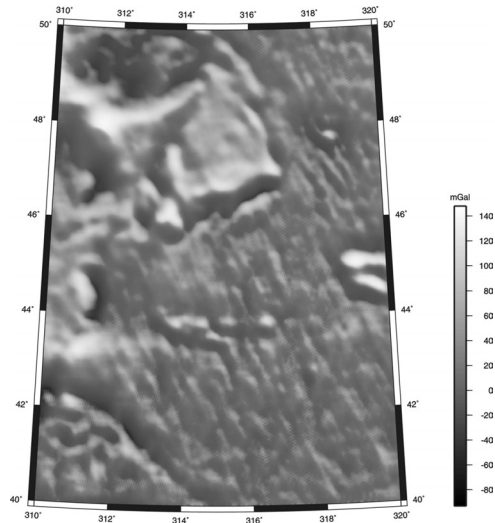


Fig. 4: Combined single-satellite gravity field in Newfoundland.

5 Validation of the Estimated Gravity Models

To assess the accuracy of the estimated gravity field models, comparisons with shipborne gravity data and the KMS01 global altimetric gravity field were carried out. Table 6 presents the comparisons between the shipborne data and the estimated gravity models. In all cases the differences were formed in the sense $\Delta g^{ship} - \Delta g^i$ where i represents the single- or multi-satellite solution. The best agreement is achieved for the GEOSAT solution, since the σ of the differences with the shipborne data is at the ± 15 mGal level. This is 0.5 and 1 mGal better than the combined and ERS1 models. Nevertheless, the differences can be considered as high since their total range is approximately 200 mGal. Comparing KMS01 with the ship data (last row of Table 6) it can be seen that the differences are almost the same, and slightly worse than what GEOSAT gives. Plotting the differences between the GEOSAT solution and the ship data (see Fig. 5) it is clear that the main part of the differences is located in the SW part of the region. Comparing the pattern of the differences in Fig. 5, with the distribution of the shipborne gravity data (cf. Fig. 2), it is clear that they are associated with a few tracks located in that region. In the rest of the area the differences are very small and present a σ of ± 5 mGal only.

Table 6. Differences between shipborne gravimetry and the estimated models. Unit: [mGal].

	max	min	mean	σ
$\Delta g^{ship} - \Delta g^{GEOSAT}$	120.2	-79.2	-0.1	± 15.5
$\Delta g^{ship} - \Delta g^{ERS1}$	118.9	-88.8	-0.1	± 16.5
$\Delta g^{ship} - \Delta g^{combined}$	120.0	-83.4	0.0	± 15.9
$\Delta g^{ship} - \Delta g^{KMS01}$	118.5	-82.6	0.1	± 15.6

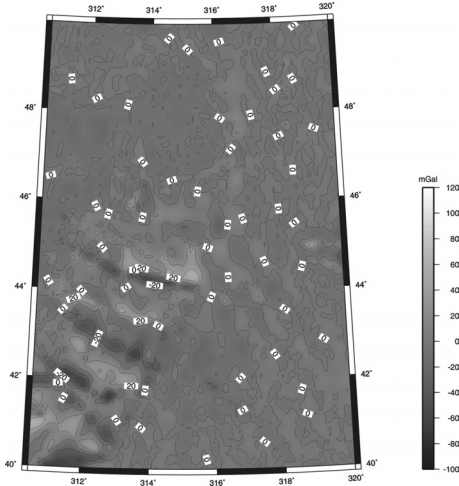


Fig. 5: Gravity anomaly differences between shipborne gravimetry and the GEOSAT solution.

Table 7 presents the gravity anomaly differences between KMS01 and the estimated gravity field models. The same conclusions as in the previous comparisons hold, since the GEOSAT solution presents the smallest differences with KMS01 (σ at the ± 5 mGal level) and the combined one gives an improved, compared to the ERS1 one, estimation of the gravity field for the area

(smaller σ of differences by 1.3 mGal). Plotting the differences between GEOSAT and KMS01 (see Fig. 6) it is evident that the differences are very small throughout the region, and noisy features are almost absent. This is an indication that we successfully managed to remove most of the SSV effects contaminating the GM altimetry data in the central-southern part of the area. The fact that the GEOSAT solution agrees better with KMS01 by 2 mGal compared to the ERS1 model, may signal that crossover adjustment is a valuable tool in the elimination of oceanic effects in the altimetric datasets, which cannot be completely removed by low-pass filtering. The latter plays an important role in the reduction of noisy features in the estimated gravity field models, but has some limitations coming from the fact that we do not know the exact spectral content of both the oceanic effects and the SSV that we wish to remove/reduce. Thus, the selection of the optimal cut-off frequency is difficult and is based on a trial and error procedure.

Table 7. Differences between KMS01 and the estimated models. Unit: [mGal].

	max	min	mean	σ
$\Delta g^{KMS01} - \Delta g^{GEOSAT}$	48.5	-34.8	-0.2	± 4.9
$\Delta g^{KMS01} - \Delta g^{ERS1}$	45.2	-33.7	-0.2	± 6.6
$\Delta g^{KMS01} - \Delta g^{combined}$	49.5	-36.1	-0.2	± 5.3

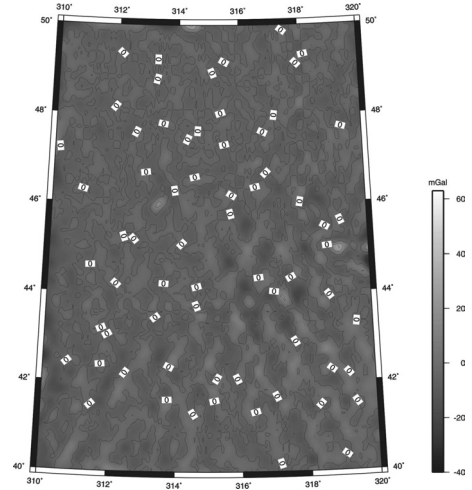


Fig. 6: Gravity anomaly differences between KMS01 and the GEOSAT solution.

6 Conclusions

Numerical investigations on marine gravity field modeling using heterogeneous data have been presented aiming at the determination of high-accuracy and high-resolution gravity solutions and the derivation of an optimal data-processing scheme for related studies. From the results and validation procedures carried out, it is evident that when altimetry data are handled properly, i.e. corrected for all error sources, blunders removed, accurate geopotential and DOT models used, the data are corrected for the QSST signal, the bathymetry is taken into account using an accurate model, the altimetry data are crossover adjusted and low-pass filtered, then, altimetric gravity field modeling accurate to

about $\pm 4 - \pm 5$ mGal (1σ) is feasible. The multi-satellite solution improves the single-satellite one from ERS1, by 1.5 mGal, in terms of the σ of differences with KMS01 gravity, while it does not manage to give better results than GEOSAT. Other studies (Andersen and Knudsen, 1998) come to the conclusion that the final accuracy is improved when combining altimetric SSHs from more than one satellites, while others (Tziavos et al. 1998) agree with our outcome. In any case, the combination of multi-satellite data sources provides the most detailed representation of the gravity field in the area, showing the importance of implementing many data sources in the solutions.

The effect of oceanic phenomena in the densely spaced GM datasets, especially in areas with high ocean dynamics, is profound and should be reduced by cross-over adjusting the altimeter datasets on a local scale as well as by low-pass filtering them. Additionally, the bathymetry should be taken into account only if an accurate model is available, whether else the accuracy of the estimated gravity model is reduced. When local bathymetry models, tested, validated and proven for their accuracy are not available, then the use of global bathymetric solutions, like JGP95E, should be implemented with caution, since they can lead to loss of accuracy. Also, the altimetry data should be corrected for the QSST signal to refer to the geoid and not the sea surface, while the question that arises is not on the necessity of such a reduction, but on the selection and the development of accurate DOT models.

Finally, taking into account that KMS01, compared to other altimetry-derived global gravity field models, e.g. the one by Sandwell and Smith, gives the smallest differences to shipborne gravimetry (see the comparisons in Andersen and Knudsen (1998)), we can conclude that the good agreement of our models with KMS01 are encouraging. This signals also the appropriateness of the methodology used, which can thus form the basis for marine gravity field modeling using satellite altimetry data. As part of our future investigations, we intend to implement not only GM but Exact Repeat Mission (ERM) data as well and take advantage of more satellites, like TOPEX/POSEIDON and JASON-1, which are not only more accurate than ERS1 and GEOSAT, but more precise as well. The combination of multi-satellite and multi-mission altimetry data can lead to a higher in resolution and accuracy gravity field model, which will lead, among others, to more accurate geoid solutions and a better understanding of marine geophysics.

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