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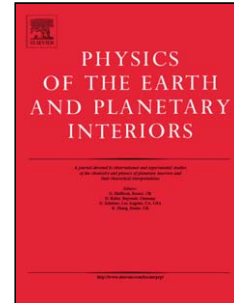
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Sensitivity analysis of the gravity geoid estimation: A case study on the Azores plateau

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Abstract

A numerical investigation into the effects of high-resolution heterogeneous data on the estimation of the geoid is presented for the area of the Azores plateau. Gravity data were complemented with high-resolution digital terrain models, mass density models from geophysical information, and the most recent geopotential model from the CHAMP mission (GGM02) to derive a high-precision gravity model. Gravity data were reduced from the residual terrain model effects and the GGM02C model, and a grid with 1.5' spatial resolution was generated by least squares collocation. The gravimetric geoid model was computed by the direct evaluation of Stokes' integral with a spherical cap radius of 0.8 degree, with gravity data reduced by Helmert's second condensation method. The geoid model was evaluated over sea by comparison with six years of Topex/Poseidon altimeter data and an absolute agreement of 14 cm RMS was achieved after removing the bias. Over land, the geoid was compared with a total of 223 GPS/levelling points corresponding to five different vertical datums. Locally adjusted solutions were determined for each island with a four-parameter model, and a maximum RMS of 12 cm (on Pico) and a minimum of 2.1 cm (on Faial) were determined. As a result of persistent volcanic and tectonic activity on most of the Azores islands, this geoid residual standard deviation must not be viewed as purely geoid error but containing, in addition, levelling and GPS position errors and vertical displacement. The effect on geoid accuracy of the digital terrain model resolution, of the spatial density of gravity data, and of the mass density was evaluated and two important conclusions were derived. First, the digital terrain model resolution has a clear effect on the geoid accuracy, with a maximum RMS of 3.4 cm on S. Miguel. Second, the mass density model does not affect the accuracy of the geoid. This second result indicates that the effect of variable density

models on geoid determination may apply only to extensive areas of high terrain and not to smaller topographic units such as single volcanoes. The high-precision gravity model and gravimetric geoid for Azores (AZGEOD07) generated in this study will allow the conversion of old orthometric heights to ellipsoidal heights at precisions useful for the monitoring of vertical deformation. The geoid will also have application to investigations of ocean circulation modulation and of dynamic processes of Earth's mantle.

Keywords: Azores; mass density, gravity; geoid;

1. Introduction

The Azores archipelago is a set of nine islands situated in the north-east Atlantic near the triple junction between Eurasia, Nubia, and North-American (Fig.1). It is a volcanic plateau with a triangular-shaped, elevated oceanic crust limited by the 2000 m isobath, crossing the Mid-Atlantic ridge and bounded to the south by the East Azores fracture zone. Although the present-day plate boundary between Eurasia and Africa appears to cut across the Azores archipelago (Lourenço *et al.*, 1998; Miranda *et al.*, 1998), the fine-scale geometry and location of this active plate boundary area remain unclear. Because of this, in the last decade much scientific attention has been placed on characterising the regional and local spatial distribution of deformation associated with the Eurasia-Nubia plate boundary (Sigmundsson *et al.*, 1995; Navarro *et al.*, 2003; Fernandes *et al.*, 2006).

The first systematic attempts to map the interplate deformation field in the area were made using the TANGO (Trans-Atlantic Network for Geodynamics and Oceanography) project (Bastos *et al.*, 1998). The initial network had one site per island and was observed periodically until 1997. In two other projects, SARAZORES (Catita *et al.*, 2005) and STAMINA (Study of the Tectonic And Magmatic Interplay in the Azores), a dense GPS network was installed on the Azores central group of islands, mainly on Terceira, Faial, and Pico. Partners in the projects have used observations from this network

to determine the intra- and inter-island displacement field over the Azores central group of islands, comprising the islands of Faial, Pico, S. Jorge, Graciosa, and Terceira.

Because available GPS records are short in duration, several attempts have been made to use conventional geodetic data (available for the whole of the 20th Century) to obtain independent estimates of the deformation field. In addition, several authors (e.g., Feigl *et al.*, 1990, Dong *et al.*, 1998) have integrated the conventional geodetic data with recent GPS observations in order to increase the temporal resolution. The high-quality record of conventional and GPS geodetic data for Faial has allowed Catalão *et al.* (2006) to demonstrate the feasibility and usefulness of the integrative method for deriving the source mechanisms for the Capelinhos eruption in 1958. These authors stressed the need for an accurate geoid surface for “planar” geodetic data reduction to the ellipsoid, rather than to sea level, both for azimuthal observations and for electronic distance measurements. For the latter, it was shown that the ellipsoidal-reduced distance measurements may change up to 7 cm, when compared with sea-level reduced measurements, thereby leading to different analysis and interpretations. The most important difference is the impact of the geoid on comparing old orthometric heights with ellipsoidal heights given by the GPS for long-term evaluation of vertical displacement or deformation. In such a case, the accuracy of the geoid surface is propagated directly to the final result and therefore a 5-7 cm accuracy level geoid surface is needed to observe secular vertical displacements and to observe the expected 10-20 cm vertical deformation (Catalão *et al.*, 2006; Sigmundsson *et al.*, 1995).

Geoid and gravity anomalies may be also used as constraints for ocean circulation modelling (Rodriguez-Velasco *et al.*, 2002; Wunsch and Stammer, 2003; Knudsen *et al.*, 2006) and for subsurface mass distribution determination (Mcadoo, 1981; Camacho *et al.*, 1997; Zhong, 2001; Kuhn, 2003). In effect, geoid and gravity anomalies reflect lateral heterogeneities in the earth’s density structure and often correlate with bathymetry, and therefore provide information concerning the subsurface density structure and the style of isostatic compensation of the bathymetric load. This

approach was used by Luis and Neves (2006) who utilised the Smith and Sandwell (1994) bathymetry grid to compute the isostatic compensation of the Azores Plateau, involving calculations of the 3D admittance between bathymetry, geoid, free-air gravity, and mantle Bouguer anomalies. At a more local scale, research efforts have been made on modelling the subsurface mass distribution associated with volcanic activity of the islands (Camacho *et al.*, 1997; Nunes *et al.*, 2006).

In the TANGO project, absolute and relative gravity campaigns were made in 1992, 1994, and 1997 (Lazaro *et al.*, 2000), aimed at determining vertical ground displacements. The coupling of geoid and surface plate motion are the principal observations that best constrain the dynamic processes occurring in Earth's mantle (Zhong, 2001). A precise geoid model, which incorporates a recent geopotential model and which uses state-of-the-art geoid computation, will allow the effects of these continental structures on the geoid to be assessed, and models of the lithosphere to be constrained.

For the Azores, some preliminary geoid solutions have been made in the last ten years (Catalão and Sevilla, 1999; Fernandes *et al.*, 2000) but have yielded rather disappointing precision, at the decimetre level, for both land and sea. The major shortcomings have been the insufficient DTM resolution and the lack of dense gravity measurements over land. However, current interest in the geoid issue has been ignited by the development of high accuracy geopotential models derived from the most recent geopotential spatial missions CHAMP, GRACE, and GGM02C (Tapley *et al.*, 2005), EIGEN (Förste *et al.*, 2005; Förste *et al.*, 2006), and the future GOCE mission (results from which should permit the geoid to be determined with unprecedented resolution and accuracy). In addition, good gravity coverage now exists over both land and sea, and a high resolution digital terrain model has been constructed, which now makes it possible to construct a high-quality gravity and geoid map for the Azores region. This present research study, therefore, focuses on the geodetic issue with emphasis on two central problems: the transformation of orthometric heights to ellipsoidal heights, and vertical datum unification. In addition to addressing these two issues concerned with geoid determination and

geoid accuracy, we investigate the effects of different terrain density values and of digital terrain model resolution and accuracy.

2. The Gravity Data Set

2.1 Land Gravity Data

Gravity data for this study were obtained from a number of prior projects and investigations. This section describes the projects and the gravity data that were generated by them, and the adjustments made to the data for the purposes of the present study.

2.1.1 Data Obtained from Previous and Ongoing Land Gravity Projects

The first gravimetric campaigns were made by in 1961 by Bettac and Scukte and the results computed by Prof. W. Grossman (in 1963) from the Geodetic Institute of the University of Hannover, as the result of a joint computation of the International Gravimetric Network which also included Lisbon station on the Portuguese mainland. At that time, a first-order gravimetric station from the International Gravimetric Network was established on S. Miguel (at the Meteorologic Observatory of Ponta Delgada), with the other satellite stations being situated at S. Miguel airport, St Maria, and Terceira (Fig. 1). Subsequently, a gravity survey for the entire Azores archipelago was made between May and August 1965, as the result of a bilateral agreement between Instituto Geográfico e Cadastral of Portugal (IGC) and the Institute Geographique Nacional (IGN) of France. During that period, 227 stations were measured using three gravimeters: two LaCoste & Romberg model G (n. 86 and 102) and one North America AG1 n. 153. On each island, a base station close to the main harbour was established as the origin of the local gravity network. Based on these base stations, local gravity networks were developed, mostly coincident with the geodetic network, with a mean density of one station per 9 km². From this campaign, 227 gravity anomalies relative to the Potsdam Gravity System (Torge, 1989) were determined according to the international gravity formula of 1930, covering all nine islands of the archipelago (Coelho, 1968).

Later, in 1977, a gravity campaign was conducted on S. Miguel by the National Institute of Meteorology and Geophysics (Mendes-Victor and Ribeiro, 1978), in which more than 800 stations were measured across the central area of the island. The base station of S. Miguel was considered to be the reference station for all measurements, and the gravity anomalies had already been computed under the Geodetic Reference System 1967 using the normal gravity formula (Heiskanen and Moritz, 1967).

In the TANGO project, absolute gravity measurements were obtained using the JILAg-5 instrument of the Finnish Geodetic Institute in 1992, 1994, and 1997 on Faial, Flores, and S. Miguel, respectively (Lazaro *et al.*, 2000). For relative gravity measurements, all nine islands were visited and 114 relative measurements were completed. The network was established in 1992 and reobserved in both 1994 and 1997. In these three campaigns, connections between the islands were also measured. Most of the sites corresponded to those of the Azores gravity network established in 1965. In 2003, all three absolute stations were reoccupied under the SARAZORES project (Mäkinen *et al.*, 2004).

Collaboration between the Instituto Astronomia y Geodesia of University Complutense and the University of Azores has led to several gravimetric studies being made across the Azores islands, with the main aim of modelling the subsurface masses distribution associated with volcanic activity. Under this project, S. Miguel, Pico, and Faial were surveyed during 1993/94, 1997, and 2000, respectively. On S. Miguel, 309 stations were measured, mainly on the east and west of the island, complementing the earlier survey of Mendes Victor and Ribeiro (1978) with the intention of covering the whole island with gravity measurements (Camacho *et al.*, 1997). On Pico, the target was the western part of the island and 250 stations were measured with a Lacoste & Romberg G933 gravimeter and using a barometer for heights. The estimated precision for the gravity was 0.02-0.03 mGal after adjustment, and 1-3 m for the heights. Faial was covered with a gravity survey of 253 stations, a density of about 1 station per km². Gravity values were referred to the absolute gravity values determined in 1997 with

JILAg-5 in the Meteorological Observatory of Horta (Faial). Heights were determined by a differential GPS survey linked to the geodetic reference network of the island and had an adopted uncertainty of 5 cm.

The most recent observations were made on Faial, in July 2006, under the TIPGRAV project, with the intention of improving the gravity geoid of the island. During the TANGO project campaigns in 1992, 1994, and 1997, relative gravity observations had already been made at nine stations. Therefore, for the 2006 campaign, it was decided to reobserve at these stations in order to compare the gravity values over time. The observations made in 1965 by the national cadastre (Instituto Geográfico e Cadastral, 1968), and by Camacho of the Instituto de Astronomía y Geodesia (Madrid) in 2000, were included in the comparison to augment the results (Bos *et al.*, 2007).

2.1.2 Adjustment of Gravity Data for this Study

In 1971, the IGSN71 standard was adopted and an offset of 14 mGal was determined between the former Potsdam Gravity System and this new standard (Torge, 1989). Bos *et al.* (2007) computed the difference with the former Potsdam Gravity System on Faial to be around 15 mGal, based on the absolute gravity measurements made in 1992, 1994, and 1997. The extra 1 mGal could be an error in the transfer of the absolute value from Potsdam to Faial. This offset of 15 mGal was applied to the gravity data acquired by Coelho (1968). All gravity data were then referred to IGSN71, and the GSR80 system was used for normal gravity. In addition, the gravity values observed in all other projects were converted to gravity anomalies using the GSR80 system.

2.2 Shipborne Gravity Data

2.2.1 Data Obtained from Previous and Ongoing Marine Gravity Projects

Marine gravity data were supplied by the NGDC (National Geophysical Data Centre, USA, GEODAS data, Version 4.1.18), covering an area with the following limits: $20^{\circ} < \varphi < 50^{\circ}$, $40^{\circ}W < \lambda < 10^{\circ}W$. Most

of these data had been acquired by institutions from the USA, United Kingdom, and France over the period 1970-2005. This data bank was improved with a recent gravimetric campaign held in 1997 under the PDIC/C/Mar project (Fernandes *et al.*, 1998) and AGMASCO project (Timmen *et al.*, 2002). The complete data set, produced by a simple merge of data files, was cleaned from repeat missions recorded in different data banks, resulting in a data set with 856141 data points and 824 tracks. All data were transferred to IGSN71 system and the anomalies converted to GRS80.

The geographical distribution of gravity data across the Atlantic is depicted in Fig. 1. In our study area, marine gravity measurements taken before the 1980s represent more than 66 % of the total of available gravity observations. Although airborne gravity data over the Azores archipelago are available, having been acquired in the AGMASCO survey, the integration of these gravity data with the marine data has already evaluated (Catalão *et al.*, 2004) and it was decided not to include these data in this study due to inexplicably high frequency errors contained in the data.

2.2.2 Adjustment of Gravity Data for this Study

The validation, adjustment, and external quality assessment of the marine gravity data was achieved using cross-over error (COE) computation and adjustment. A dedicated software package, VALDAMA, was written for this purpose (Catalão and Sevilla, 2004). VALDAMA determines the intersections between tracks, computes the differences between the estimated anomaly in the intersection point from both tracks and, finally, adjusts the residuals, estimating the best bias for each track. Track bias was determined through a global adjustment of the external COE and different weights were applied to each observation equation. The minimum constraint adjustment was adopted with a one-step global adjustment, constraining to a zero bias, of the PDIC/C/MAR observations surveyed in the Azores in 1997. A standard deviation was attributed to each cruise or leg based on its surveying date. For the most recent tracks (post-1980) a 2.5 mGal standard deviation was used, for the tracks between 1970 and 1980 a value of 7.5 mGal was used, and for tracks prior to 1970 a 12.5 mGal standard deviation was applied.

After the adjustment, the standard deviation of the cross-over errors was reduced to 3.8 mGal from an initial RMS of 12.3 mGal, with a minimum and maximum reduced to -29 and 47 mGal, respectively. The adjustment solution (bias) was applied to each track and the associated estimated standard deviation for an individual track was assigned to each observation belonging to the track. The resulting file contains 533718 gravity anomaly observations each with an associated standard deviation that does not reflect an observation's internal precision but rather the overall quality of the individual track to which it belongs.

2.3 Satellite Altimeter-Derived Gravity Grid

There are several public domain, altimeter-derived, gravity anomaly data sets with global coverage. These models, in grid format, have been computed directly from sea surface heights, from KMS02 (Andersen *et al.*, 1999), or through deflections of the vertical (Sandwell and Smith, 1997). KMS02 is the latest of a series of satellite altimeter-derived gravity anomaly data sets supplied by Kort-og Matrikel Stryrelsen, and is available as a download, currently via <ftp.kms.dk>. Numerous comparisons between marine observations and satellite altimeter-derived gravity anomaly data sets have been presented (e.g., Sandwell and Smith, 1997; Andersen and Knudsen, 1998). The reported precision ranges from 3 mGal to 14 mGal, varying as a function of location. In the North Atlantic, the reported precision for the Sandwell and Smith (1997) model is 7.6 mGal and for the KMS model is 5.8 mGal (Andersen and Knudsen, 1998). KMS02 exhibits the best agreement with ship-borne gravity data and because of this we have adopted this model (KMS02) as our background model. KMS02, with a resolution of 0.033° , covers the entire marine area and also overruns land areas such as the Azores islands. Because of altimeter gravity data degradation around coastal regions, a mask with respect to each coastal line, with an offshore buffer of 20 km, was constructed and used to clean all satellite data inside those areas.

3. The Geopotential Model

With the recent dedicated gravimetric satellite missions CHAMP and GRACE, and the in-coming GOCE mission, the long wavelengths (about 200 km) of the gravity field will be accurately determined just with satellite data. Based on GRACE data, a new generation of Earth geopotential gravity field models have been derived, namely EIGEN-GL04C (Förste *et al.*, 2005, 2006) and GGM02C (Tapley *et al.*, 2005). EIGEN-GL04C is a combination of GRACE mission and Lageos mission data with the inclusion of altimetry and gravity surface data. The model is complete to degree and order 360 in terms of spherical harmonic coefficients, and resolves wavelengths of 110 km in the geoid and gravity anomaly fields. The GGM02S geopotential model is determined only from GRACE data spanning from April 2002 to December 2003, and includes no constraints. From the combination of this model with terrestrial gravity information (surface gravity and mean sea surface), the higher resolution global gravity model GGM02C was developed, complete to degree and order 200. The higher degrees were constrained to the harmonic coefficients of geopotential model EGM96 (Lemoine *et al.*, 1997) and the (2,0) harmonic was constrained to its long-term mean value using EGM96. In this study, the GGM02C original model, complete to degree and order 200, was extended to degree and order 360 by using the EGM96 coefficients.

The C(2,0) coefficient of the GGM02C geopotential model is a zero tide value accordant with the IAG resolution n. 16 of 1984. This means that the resulting geoid model is a zero tide system. Different permanent tide systems will produce different geoid models relative to a fixed ellipsoid, differing in shape up to 5 cm as a function of latitude (Basic and Rapp, 1992). The difference between the geoid undulations computed on the zero tide and the tide free systems decreases monotonically from the equator, with a value of 9 cm at the poles and crossing the zero value at approximately 35° latitude. It means that, at the latitude of the Azores (38°), the difference between geoid solutions in zero-tide and free-tide systems is negligible (about 3 mm). Most of our height data are on a free tide system including GPS ellipsoidal heights, orthometric heights, and sea surface heights.

The gravity anomaly is computed from the global geopotential model using:

$$\Delta g_{\text{GGM}}(\varphi, \lambda, r) = \frac{GM}{r^2} \sum_{n=2}^{n_{\text{max}}} (n-1) \left(\frac{a}{r}\right)^n \sum_{m=0}^n (\Delta C_{nm} \cos m\lambda + \Delta S_{nm} \sin m\lambda) P_{nm}(\sin \varphi) \quad (1)$$

and the geoid height is:

$$N_{\text{GGM}}(\varphi, \lambda, r) = \frac{GM}{r\gamma} \sum_{n=2}^{n_{\text{max}}} \left(\frac{a}{r}\right)^n \sum_{m=0}^n (\Delta C_{nm} \cos m\lambda + \Delta S_{nm} \sin m\lambda) P_{nm}(\sin \varphi) \quad (2)$$

where: (φ, λ, r) are the spherical (geocentric) coordinates of point P; ΔC_{nm} and ΔS_{nm} are the fully normalized geopotential coefficients of the anomalous potential; GM is the geocentric gravitational constant; a is the semi-major axis of the reference ellipsoid; P_{nm} are the fully normalized Legendre functions; γ is the normal gravity (GRS80) at point P; and n_{max} is the maximum degree of the geopotential model.

The reduced gravity anomalies Δg_{red} are obtained by subtracting the free-air gravity anomalies from the geopotential model anomalies:

$$\Delta g_{\text{red}} = \Delta g_{\text{FA}} - \Delta g_{\text{GGM}} \quad (3)$$

If the gravity anomalies determined by the global geopotential model are close to the observed gravity anomalies, it is reasonable to assume that the geopotential model is suitable to represent the long wavelength for a regional geoid model. Following this reasoning, and having at the time at least two new geopotential models derived from GRACE data, the three geopotential models (namely, EIGEN-

GL04C, GGM02C, and EGM96) were evaluated by computing and analysing the resulting reduced anomalies. The results are presented further below and include the residual terrain effects.

4. The Digital Terrain Model and Topo-Bathymetric Effects

A digital terrain model covering the Azores area was constructed (AZDTM) by compiling and reprocessing medium-scale terrestrial maps (1:25,000) and ship track data with single beam and multi-beam acquisition (around the Azores archipelago). In order to accommodate different needs and different scale maps, two different resolutions were adopted for the DTM. For the land and surrounding areas, a detailed grid was chosen with a resolution of $0.001^\circ \times 0.001^\circ$ (about 100 m), with the limits 29.3W to 24.5W and 36.4N to 39.6N (Fig. 2). For the distant topographic masses, a coarse grid with a resolution of 0.005° was used, with the limits 34W to 20W and 33N to 45N. The resolutions used are multiples of each other and are both integer multiples of one degree.

The classical terrain correction, in planar coordinates and for a constant density ρ , is given by (Forsberg, 1984):

$$C_p = G\rho \int_{x_1}^{x_2} \int_{y_1}^{y_2} \int_{z_1}^{z_2} \frac{(z - h_p)}{r^3} dx dy dz \quad (4)$$

where: G is the gravitational constant; ρ is the mass density; r is the distance between the computation point (x_p, y_p, h_p) and the integration point (x, y, z) . The vertical integration is taken from the geoid, z_1 , to the surface, z_2 . The integral is evaluated over the irregularities of the topographic mass relative to a Bouguer plate passing through the computation point. The terrain correction was computed for all individual data points by means of the prism integration procedure implemented in the TC program of the GRAVSOFTE package (Tscherning *et al.*, 1994). This terrain correction was added to the reduced anomalies in order to obtain the fully reduced anomalies, in agreement with Helmert's second condensation method.

The process of shifting masses that lie above the geoid to a surface layer on the geoid in the Helmert condensation method implies, in practice, a gravity reduction and hence a change in the gravity potential and the geoid. This change is referred as the indirect effect. The indirect effect of Helmert's second condensation reduction on the geoid is given in planar approximation by (Wichiencharoen, 1982):

$$N_{\text{ind}} = -\frac{\pi G \rho}{\gamma} h^2(x_p, y_p) - \frac{G \rho}{6\gamma} \iint \frac{h^3(x, y) - h^3(x_p, y_p)}{s^3} dx dy \quad (5)$$

where: s^3 is the planar distance between the point P and the integration point. The indirect effect was computed for all the Azores islands and a maximum of -0.17 m was obtained for the volcano summit of Pico at an altitude of 2430 m.

In order to compute the terrain correction and the indirect effect, a mean terrain density ρ must be selected. This density value is of crucial importance and may affect geoid solutions. It is therefore important to base the selection of the local or regional terrain density on constraints based either on numerical considerations or on direct measurements. For the Azores, experimental density determinations indicate values ranging from 2250 kg m⁻³ to 2840 kg m⁻³ for lava flows of basaltic composition (Motta and Nunes, 2003). Besides this direct determination, other authors using indirect methods such as Nettleton's or Camacho's method (Camacho *et al.*, 2002) have used different density values for each island of the Azores for the purposes of gravity data inversion. Camacho *et al.* (1997) chose a value of 2300 kg m⁻³ for the mean density for S. Miguel, Montesinos *et al.* (2003) proposed a density of 2480 kg m⁻³ for Terceira, Nunes *et al.* (2006) used a mean value of 2520 kg m⁻³ for Pico, and Camacho *et al.* (2006) adopted a value of 2390 kg m⁻³ for Faial. These four islands represent more than 80 % of the topographic area of the Azores islands and a representative mean density value may be obtained by averaging the four density values, resulting in a mean value of 2423 kg m⁻³. A water density of 1000 kg m⁻³ for terrain correction of the bathymetry was used.

Because the mean terrain density adopted (2423 kg m^{-3}) is less than the mean density of the upper lithosphere (2670 kg m^{-3}), the effect of this difference on the terrain correction and the indirect effect was analyzed. The maximum difference between the two computations using the different density values for the terrain correction is only 6.3 mGal with a standard deviation of 0.09 mGal. The effect of different mass densities on the computed indirect effect is up to 2 cm. A simplified approximation of the difference between the two different density models on the gravity quantities may be given by a Bouguer plate with a height given by the standard deviation of the mean height of all islands. This corresponds to the expression $2\pi G(\rho_0 - \rho_1)H_{\text{mean}} = 0.0103 * H_{\text{mean}} \text{ mGal}$. The mean height of the Azores islands determined by the AZORDTM model is 330 m and thus a rough value of 3.3 mGal for the difference between two density values is expected. The effect of these small differentials between different density values on the precision of the geoid is analyzed further below.

5. Gravity Data Merging and Gridding

In this project, three main sources of gravity measurements were used: land-based gravity data; shipborne gravity data; and satellite altimeter-derived gravity data. Prior to their use, gravity data must be both internally and externally validated and merged into a unique and consistent gravity dataset amenable to further processing and analysis, particularly for geoid determination. There are essentially two main approaches for combining multi-source gravity data: one operates on the space domain working at the observation level, known as the draping technique (Strykowski and Forsberg, 1998); and the other operates on the frequency domain, combining data spectra (Kern *et al.*, 2003). In this paper, we applied an alternative space domain method based on least squares optimal interpolation (Catalão, 2006) that combines two or more gravity data sets with their associated error variances into a resulting gravity data set that is statistically optimal. The merging process was performed in three steps: first, land data were analyzed for temporal variations and for bias detection between surveys; second, land data were merged with satellite-derived gravity anomalies and a background grid was

derived; and third, the shipborne data were assimilated with the background model using least squares optimal interpolation.

For the first step, it was appreciated that land data had been acquired by different institutions over different periods spanning more than 40 years. The accuracy that should be attributed to these gravity anomalies is mainly a function of the measurement errors, the accuracy of the gravimeter, and the height error. For the latter, the barometric levelling used to determine the height of some stations may lead to errors of 1-2 mGal in the free-air reduction. In order to better estimate the sum of all observational noise and the accuracy of the previous surveys (from 1965 to 2000), a comparison with recent surveys, linked to the absolute stations of S. Miguel and Faial, was made. In addition, repeated absolute and relative measurements, although limited to Faial, would provide a good measure of the misfit between surveys and a first estimate of the final accuracy that should be attributed to the computed gravity.

Relative gravity data were acquired in 1965 by Coelho (1968), in 1977 by Mendes-Victor and Ribeiro (1978), and in the period 1998-2002 by Camacho *et al.* (2007). To determine possible biases between surveys and to evaluate the accuracy of the former ones, repeated surveys were analyzed for S. Miguel, Terceira, Pico, and Faial. For Faial, nine stations were observed in 1965, 1997, and 2006 (Bos *et al.*, 2007), and it was verified that the largest temporal changes are around 0.2 mGal. The same level of temporal variation was determined for the other three islands and a prudent repeatability of 0.3-0.4 mGal was assumed. Again on Faial, where a dense gravity network exists, a comparison between 40 points observed both during the 2000 and 2006 campaigns demonstrated that the agreement for 24 stations was within ± 0.05 mGal. For the remaining 16 stations the larger discrepancies could only be partially attributed to differences in location. Therefore, it seems prudent to assign the likely repeatability of the gravity observations, surveyed in both 2000 and 2006, at ± 0.1 mGal. This represents a very good agreement between these two data sets and accordingly a 0.1 mGal standard error was assigned to the Faial gravity measurements. For the other islands, a standard error of 0.4

mGal was assigned to the gravity measurements. This level of agreement between surveys separated by some 40 years incorporates not only measurement errors but also vertical deformation due to volcanic activity on these islands. However, this result, although good enough for geoid determination, is unsuitable for vertical land deformation studies where μGal level uncertainties are required. Overall, a total of 2074 gravity measurements were validated and linked to Faial and S. Miguel absolute stations.

For the second step, the background model was constructed from satellite-derived gravity anomalies of the KMS02 gravity data model. Because of altimeter gravity data degradation along coastal regions, a mask with respect to each coastal line, with an offshore buffer of 20 km, was constructed and used to clean all satellite data inside those areas. Land data on the Azores islands was then used to fill these cleaned areas. The final grid therefore comprised satellite-derived gravity anomalies over sea and observed anomalies over land areas, with the same correlation distance of 10 km as used to compute the KMS02 model. The process of gridding is subject to aliasing in the presence of a high-frequency signal. With gravity data, part of this high frequency signal is due to the topography and can be removed from the original signal by removing the residual terrain model effect on the Bouguer plate. The resulting residual anomalies are given by:

$$\Delta g_{RTM} = \Delta g_{FA} - \Delta g_{GGM} - 2\pi G\rho (H_p - H_{ref}) + C_p \quad (6)$$

The gravimetric terrain correction C_p was calculated from the irregularities of the topography, after removal of a Bouguer plate. The term $2\pi G\rho(H-H_{ref})$ was estimated from the irregularities of the topography with respect to a topographic reference surface of 100 km resolution coinciding with the resolution of the geopotential model. The reference surface was generated by low-pass filtering the original 100 m resolution DTM. For heights above sea level, a ρ value of 2670 kg m^{-3} was used and for heights below sea level a density of 1000 kg m^{-3} was adopted. The standard deviation of the

residual gravity anomalies is reduced from 23 mGal to 9.4 mGal when the GGM02C geopotential model (augmented to degree 360 using EGM96) is used, and the mean value is reduced to zero (Table 1).

The gridding was performed using least squares collocation. Data were detrended prior to gridding by removing the mean value and using a Gauss Markov covariance function model (an exponential function that decays with radial distance) with a correlation length of 20 km. The correlation length was determined from the empirical covariance function computed with the residual gravity anomalies. The computation point was the centre of a cell with a resolution of 0.01 degrees (about 100 m), extending from 35.505 to 41.495 in latitude and from 32.495W to 24.005W in longitude, and containing 600 rows and 850 columns.

The third and final step involved the merging of shipborne data with the background gravity model. The discrepancies between the background model and the shipborne measurements were computed and exhibit a mean value of -0.96 mGal and a standard deviation of 7.1 mGal in agreement with the results presented for this area by Andersen and Knudsen (1998). Prior to the merging process the bias must be removed and the biased data set, or data sets, must be identified. Since marine gravity data were correctly adjusted into a coherent reference system by a minimum constraint (given by the most recent cruise) the bias is removed by adding 0.96 mGal to all marine data. This is equivalent to consider that the reference cruise was affected by a tie-in error of 0.96 mGal that was not considered in the adjustment.

The unbiased residuals, associated background error covariance, and observation error model covariance were used to derive a correction surface by least squares optimal interpolation. The correction surface was assimilated into the background model constructed in step 2, yielding the best possible estimate of the gravity field. Further details about the merging process can be found in Catalão (2006). The resulting gravity model exhibits a misfit against the marine observations of 2.6 mGal, with considerable recovery of the high frequency spectrum of the gravity field. The

geopotential model contribution was then repositioned and the RTM topo-bathymetric effects were added (Fig. 3). It should be noted that the removal of both effects was conducted point-wise and that the restoring was conducted also point-wise but in grid format.

Analysis of the bathymetric map (Fig. 2) shows a succession of basins and islands that were jointly called Terceira Rift, presenting a characteristic distance of 20 to 30 km from S. Miguel to the Middle-Atlantic Ridge (MAR). The prominent basins are Hirondele (south and north) and Graciosa (west and east). To the south is the well-defined East Azores fracture zone delimiting the Azores plateau and the MAR is seen oriented N10E. Free-air gravity anomalies (Fig. 3) over the MAR, along the axial valley, are negative with amplitudes of several tenths of mGal, and with peak-to-peak horizontal distances of approximately 50 km, which is generally the case with slow-spreading ridges. The 50 km peak-to-peak pattern is more pronounced on the gravity map than on the bathymetric map. The gravity anomaly map is strongly correlated with the bathymetric map, where the islands are visible as large positive gravity anomalies and basins correspond to large negative gravity anomalies. The terrain-corrected residual anomalies (Eq. 6) are as a result very smooth and this explains the good results of the least squares optimal interpolation scheme. Geophysical interpretations of the bathymetry and of the gravity anomalies have been made by Lourenço *et al.* (1998) and Luis *et al.* (1998), respectively.

With the aim of evaluating the best geopotential model for this area, residual gravity anomalies were computed with the GGM02C, EIGEN-GL04C, and EGM96 geopotential models, and RTM effects were also removed. The results (Table 1) show that using the EIGEN-GL04C geopotential model produces a larger variance of the residuals than the other two models, with a small difference of less than 1 mGal on the standard deviation. Although these two new geopotential models incorporate data from the most recent satellite missions (GRACE and CHAMP), they are not superior to the EGM96 model, at least in this study area. The gravimetric effect of the geopotential and residual terrain models is pronounced, reducing the gravity field variability almost six-fold, from 529 mGal² to 88 mGal². This result has enhanced the high quality of the DTM used for the RTM computation. The result also

represents a remarkable improvement on earlier solutions (Catalão and Sevilla, 1999; Fernandes *et al.*, 2000) in which residual standard deviations of 12 and 19 mGal were obtained, respectively. Based on the results presented in Table 1, model GGM02C was adopted as the reference model for the study.

Insert Table 1.

6. Determination of the Gravimetric Geoid

Stokes' integral for computing the geoid undulation N at an arbitrary point on the geoid can be expressed as (Heiskanen and Moritz, 1967):

$$N(\phi, \lambda) = \frac{R}{4\pi\gamma} \iint_{\sigma} S(\psi) \Delta g(\phi', \lambda') d\sigma \quad (7)$$

where: Δg are the gravity anomalies; $S(\psi)$ is Stokes' function, σ is Earth's surface; R is the mean Earth radius; and γ is the mean normal gravity. There are two drawbacks concerning the direct application of this expression. First, the integral should be evaluated for the whole Earth and, second, there should be no masses on the exterior of the geoid. The former problem can be solved by the use of a global geopotential model removing the long-wavelength part of the signal from the gravity anomalies, thereby regionalizing the signal. The latter problem implies that the gravity measurements must be reduced and the above-geoid masses removed. This may be done using one of several available methods (Heiskanen and Moritz, 1967). In this paper, masses were shifted to the geoid using Helmert's second condensation reduction. Helmert's second method of condensation, and the associated downward shift of the data from the surface of the topography to the geoid, is given approximately by the terrain correction, C_p . A grid of Faye (free-air) anomalies is generated as an approximation to Helmert anomalies.

The resulting residual gravity anomalies are obtained by:

$$\Delta g_{res} = \Delta g_{FA} - \Delta g_{GGM} + C_p \quad (8)$$

where: Δg_{res} is the residual gravity anomaly; Δg_{FA} is the observed free-air anomaly; Δg_{GGM} is the gravity anomaly derived from the geopotential model GGM02C; and C_p is the gravimetric terrain correction. The statistics for these residual gravity anomalies calculated are presented in the final row of Table 1. Stokes' integral was applied to the residual anomalies, after which the low frequencies were restored as geoid undulation given by the earth geopotential model and the high frequencies as the indirect effect, due to terrain correction. The final geoid undulation is given by:

$$N = N_{GGM02C} + \frac{R}{4\pi\gamma} \iint_{\sigma_0} S(\psi) (\Delta g - \Delta g_{GGM02C} + C_p) d\sigma + N_{ind} \quad (9)$$

where: N_{GGM02C} is the geoid undulation derived from the GGM02C geopotential model; and N_{ind} is the indirect effect (Wichiencharoen, 1982) as a result of using Helmert gravity anomalies in Stokes' formula. The long-wavelength and short-wavelength contributions have already been computed in Sections 3 and 4 above.

The evaluation of Stokes' integral can be achieved using several methods. One of the most common is the Fast Fourier Transformation (FFT) technique in its various sub-strategies including the Fast Hartley Transform (Tziavos, 1993), the 2-D spherical multiband FFT (Forsberg and Sideris, 1993), and the spherical 1-D FFT (Haagmans *et al.*, 1993). Tziavos *et al.* (1988) demonstrated that, in general, these techniques do not present numerical instabilities (Tziavos *et al.* 1988) and present identical results to methods using analytical evaluations (numerical integration) of Stokes' integral (Li and Sideris 1994). Nevertheless, due to the small geographical extent of the study area, the geoid was

computed by numerical integration of Stokes' integral truncated to a limited region σ_0 given by the spherical distance ψ_0 around the computation point. Some preliminary tests were performed in order to determine the appropriate spherical distance, and a value of 0.8 degree was identified. This value was determined from a preliminary fast external evaluation of the geoid model computed with different spherical distances from 0.5 to 2.0 degrees.

Following the methodology as described, several evaluation tests were performed with the digital terrain model resolution, the gravity data density, and the mass density. Three geoid solutions were computed. Two geoid solutions with a mass density of 2670 kg m^{-3} and different digital terrain models with 500 m and 100 m planar resolutions respectively, and a third solution with a mass density of 2423 kg m^{-3} and a digital terrain model with 100 m resolution. The geoid model for the standard mass density and 100 m DTM resolution is depicted in Fig. 4. In the area of investigation, the geoid undulation is dominated by very long wavelength variation of about 11 m in the N-S direction, being produced mainly by the global geopotential model. On top of this pattern, shorter wavelength variations are discernible, with an amplitude of about 1 m in the WNW-SSE direction.

7. Results and Discussion

The purpose of this study was to determine a high accuracy geoid model that will serve as a geodetic reference surface for vertical datum unification between the various islands of the Azores, and which will enable the transformation of GPS to orthometric heights within each island. Temporal vertical deformation can now be determined for more than 60 years within the expected sub-decimetre accuracy. In practice, this can be achieved by applying the simple relationship between the two height types for each geodetic vertex (Heiskanen and Moritz, 1967):

$$h = H + N \quad (10)$$

where: h is the ellipsoidal height obtained from GPS; H is the Helmert orthometric height obtained on the Azores from trigonometric levelling; and N is the geoid height above the ellipsoid given by the computed geoid model. Over sea, the equivalent relation should be given by:

$$\text{MSSH} = N + \text{MDT} \quad (11)$$

where: MSSH is the ellipsoidal mean sea surface height; and MDT is the mean dynamic topography.

Due to several factors, including datum inconsistencies inherent to each vertical system that refers to a different reference surface, or systematic effects and distortions in the height data (ellipsoidal and orthometric), or long-wavelength geoid errors, the above expressions are not exact and must not be directly applied. Instead, a correction surface model must be used to account for systematic effects and datum inconsistencies. The most common model used for this purpose is the simplified polynomial transformation given by

$$(h-H)-N = N_{\text{geom}} - N_{\text{grav}} = f(\phi, \lambda) + v_{ij} \quad (12)$$

where: $h-H-N$ can be substituted by $\text{MSSH}-\text{MDT}-N$ over sea; and $f(\phi, \lambda)$ is usually a two-dimensional polynomial function of n parameters obtained from the resolution of this equation by least squares. A fourth-order polynomial parameter model is frequently used, absorbing most of the datum inconsistencies among the two height data systems and also long-wavelength geoid errors. The latter are considerably reduced with the new, augmented GGM02C geopotential model. Higher-order correction surfaces will lead to a better adjustment of geoid and ellipsoidal height but will also absorb GPS and network error into the model. The most common four-parameter model used is the simplified model given by (Heiskanen and Moritz, 1967):

$$N_{\text{geom}} - N_{\text{grav}} + v_{ij} = \begin{bmatrix} \cos\varphi \cos\lambda \\ \cos\varphi \sin\lambda \\ \sin\varphi \\ 1 \end{bmatrix} \begin{bmatrix} \Delta x \\ \Delta y \\ \Delta z \\ N_0 \end{bmatrix} \quad (13)$$

where: Δx , Δy , Δz are the datum shift parameters; and N_0 is the geoid bias relative to the mean sea height given by Topex altimeter data.

The external evaluation of the accuracy of the geoid model was achieved by adjusting the geoid model to the mean sea surface obtained from Topex/Poseidon altimeter data and then comparing the resulting geoid model to a set of geodetic vertices where GPS and orthometric heights exist. The mean sea surface was supplied by Calvão and Sevilla (2006) and comprises Topex/Poseidon altimeter data from cycle 3 to cycle 235 representing the mean sea surface of the last six years. The mean dynamic topography was supplied by Rio and Hernandez (2004) and was determined from oceanographic data. It can be assumed that this mean surface, sea surface height plus the mean dynamic topography, is a good approximation of the actual geoid equipotential surface. Adjusting the geoid model to this mean sea surface data using the four-parameter transformation results in a further geoid model with a W_0 value close to that determined by Bursa *et al.* (2002). The results of this process (Table 2) are reported for the geoid solution computed with 100 m DTM resolution and 2670 kg m^{-3} mass density (see Fig. 4).

Insert Table 2.

The statistics reported in Table 2 reveal that there is no improvement to the geoid model accuracy after the adjustment using Topex/Poseidon altimeter data. This means that there is no significant tilt or datum shift between altimeter data and the geoid, but a considerable bias of -0.917 m was detected and recovered into the adjusted geoid model. Geoid precision is on the 14 cm level over sea (Table 2), which is a typical value for Topex/Poseidon altimeter data and geoid comparisons (Vergos *et al.*, 2005; Knudsen *et al.*, 2006).

Over land, a total of 223 geodetic vertices with known orthometric and ellipsoidal height were used for evaluating the geoid. The orthometric heights were determined between 1940 and 1950 across the whole archipelago and reobserved on Faial in 1987 and 1997. The orthometric heights are referred to the local vertical datum, one for each island. The ellipsoidal heights (h) refer to the ITRF94 reference frame and were surveyed between 1995 and 2000.

The mean residual between the gravimetric geoid and the GPS/levelling vertices varies from 0.65 m on S. Miguel to 0.12 m on Pico. These mean residual values may be interpreted as the vertical datum shift relative to the present-day mean sea surface, and a relative bias between local vertical data may be assessed. However, a better and more realistic value may be assessed if the geoid residual for the tide gauge is known. Although such a precise reference does not exist, there are three continuous GPS stations close to the local vertical datum: on S. Miguel (PDEL, IGS station), Terceira (TERC), and Faial (FAIM). At these three sites the residual geoid is 0.723 m, 0.281 m, and 0.152 m, respectively. There is an almost perfect agreement between these residual values and the mean residual for each island. It seems that these sites are likely to represent each island's vertical datum shift. In addition, the mean residuals are all positive, indicating an overall sea level rise and/or land subsidence. A sea level rise value of 1.3 mm/year determined by Dias and Taborda (1992) would have a direct effect of about 8 cm on the geoid residual and, assuming that the sea surface topography is constant across the Azores, a relative maximum vertical datum shift of 53 cm is determined, between Pico and S. Miguel. This large relative bias could be explained by differences in volcanic-tectonic activity on these islands or by incorrect tide gauge measurements.

The standard deviation was computed for each island and ranged between 4.2 cm on Faial and 18.3 cm on Pico. Although this is the best way to assess geoid precision, in this particular case (the Azores) the geoid residual standard deviation must not be seen as purely geoid error but containing, in addition,

levelling and GPS position errors and vertical displacement. At a regional scale, the results of Table 2 provide the expected precisions for transforming orthometric to GPS heights.

The other two geoid solutions (DTM 500m / $\rho=2670 \text{ kg m}^{-3}$, DTM 100m / $\rho=2423 \text{ kg m}^{-3}$) were determined to evaluate the effects of digital terrain model resolution (and implicit precision) and of the mass density model on the precision of the geoid. The results of these two geoid models are presented in Table 3, and are compared with the previous solution (DTM 100m / $\rho=2670 \text{ kg m}^{-3}$). Two important conclusions derive from these results: first, the digital terrain model resolution has a clear effect on the precision of the geoid; and second, the mass density model does not affect the geoid precision. The increase in the resolution of the digital terrain model (from 500 m to 100 m) has the consequence of increasing the precision of the model heights, reflected directly in the computed gravimetric terrain corrections. The improvement is a maximum of 3.4 cm on S. Miguel and a minimum of 0.5 cm on Faial. Regarding the mass density variation, the results are somewhat unexpected, indicating no improvement in geoid accuracy when a local density model (2423 kg m^{-3}) is considered rather than the global mean value of 2670 kg m^{-3} . Similar results have been produced by Tziavos and Featherstone (2000), in even more complicated lithospheric structures in south-west Australia. The explanation may be that, although the mass density enters directly into the gravimetric terrain corrections, its effect will be noticeable only for high altitudes, which is not the case of the Azores islands at a mean altitude of 330 m. An improvement of 3 mm was detected between the density models for the highest summit caldera of the Azores (2410 m) on Pico, but was insufficient to affect the overall result for the island.

Insert Table 3

Finally, a locally adjusted solution was determined for each island with a four-parameter model (Table 4). This geoid solution will be useful for local transformations between orthometric and GPS heights. The geoid residual standard deviation was reduced to 12 cm on Pico and to 2.1 cm on Faial. The result for Faial is very promising, and will enable old orthometric heights to be transformed to ellipsoidal heights for the determination of precise vertical deformations on that island. The same reasoning applies to the other islands but with less precision.

Insert Table 4

Each island has a different vertical datum and has experienced different tectonic/volcanic activity during the period starting from the trigonometric survey (1940-1950) to the GPS survey (1955-1997). Therefore, on these dynamic islands, the residuals between the gravimetric and geometric geoids may be due to local vertical deformation in addition to the expected observational and modelling errors. For example, in 1997, simultaneous trigonometric and GPS surveys were performed on Faial, during which pairs of ellipsoidal and orthometric heights were determined. This means that for a sub-set of geodetic vertices (10 vertices) there are almost simultaneous ellipsoidal and orthometric height differences. There are 26 height differences between the 10 vertices, hence a relative evaluation of the accuracy of the geoid is possible. The difference in the geoid undulation between two vertices can be approximated by the differential between the ellipsoidal and orthometric height differences at the two vertices:

$$dh - dH - dN = v_{ij} \quad (14)$$

and the four-parameter transformations used above can be converted into:

$$dN_{\text{geom}} - dN_{\text{grav}} + v_{ij} = \begin{bmatrix} -\sin\phi \cos\lambda d\phi - \cos\phi \sin\lambda d\lambda \\ -\sin\phi \sin\lambda d\phi - \cos\phi \cos\lambda d\lambda \\ \cos\phi d\phi \end{bmatrix} \begin{bmatrix} \Delta x \\ \Delta y \\ \Delta z \end{bmatrix} \quad (15)$$

in which the four-parameter transformation now becomes a three-parameter transformation because the zero degree term is eliminated due to the differencing. Because of this, the bias between the geoid and the local vertical datum cannot be determined. A system of 26 equations was solved for three parameters and a condition equation was included forcing the geoid model to assume the value of 58,932m at the continuous GPS station FAIM on Faial. The residuals for the 32 geodetic vertices were computed and a standard deviation of 1.7 cm was found. Over Faial, the geoid model performs much better than on other islands, partly because of the high quality and density of gravity measurements, about one measurement point per km², but mainly because of the recent orthometric heights used in the external evaluation. The external evaluations performed for the other islands include not only the geoid errors but also the old orthometric errors and possible temporal vertical deformation.

8. Summary and Conclusion

A new, high-precision geoid model, AZGEOD07, has been computed for the Azores archipelago and surrounding marine area. This model incorporates the most recently released geopotential model (GGM02C from the GRACE mission), a high resolution digital terrain model, and dense land gravity data acquired in the last decade.

Land gravity data were validated and partially reprocessed to accommodate the recent absolute measurements made on S. Miguel and Faial. Shipborne data were crossover adjusted and then merged with the KMS02 gravity model using least squares optimal interpolation. Long and short wavelengths given by the geopotential model and the AZDTM100 digital terrain model were removed from the gravity data, having first computed a grid of 0.025 degree resolution to avoid aliasing effects.

Removed spectra were afterwards restored and a final free-air gravity model, AZGRAV07, was computed.

Different geopotential model solutions were evaluated and it was found that the GGM02C geopotential model, augmented to degree 360 using EGM96, fits the local gravity field slightly better than other geopotential models with a small difference of less than 1 mGal on the standard deviation. A new digital terrain model reduced the gravity field variability almost six-fold from 529 mGal² to 88 mGal². This is a remarkable improvement on earlier solutions, and reflects the high quality of the DTM used in the RTM computation.

The geoid was computed by the direct evaluation of Stokes' integral with a spherical cap radius of 0.8 degrees with gravity data reduced by the Helmert's second condensation method. The long wavelengths from the geopotential model were subsequently repositioned and the indirect effect added.

Geoid accuracy was evaluated both over sea and over land. The marine evaluation was achieved by comparing the geoid with six years of Topex/Poseidon altimeter data. An absolute agreement of 14 cm RMS was determined, the systematic datum difference having been removed. Over land, a total of 223 geodetic vertices with known orthometric and ellipsoidal heights were used for the geoid evaluation. The mean residual between the gravimetric geoid and the GPS/levelling vertices was determined for each of the nine islands, and interpreted as the vertical datum shift relative to the present-day mean sea surface. The relative bias between local vertical data was assessed and a maximum relative vertical datum shift, of 53 cm, was determined between Pico and S. Miguel. This large relative bias could be explained by differences in volcanic/tectonic activity between these islands, or by incorrect tide gauge measurements.

Using a four-parameter model, a locally adjusted solution was determined for each island. The geoid residual standard deviation was considerably reduced, to 12cm for Pico and to 2.1 cm for Faial. The locally adjusted solutions mean that old orthometric heights will be able to be transformed to ellipsoidal heights, allowing vertical deformations to be determined for all islands, most precisely on Faial.

The effects of the digital terrain model resolution, of the spatial density of gravity data, and of the mass density on the accuracy of the geoid were evaluated by computing three geoid solutions. Comparisons between the geoids showed that the digital terrain model resolution has a marked effect on the accuracy of the geoid, but that mass density differences do not. In this study, real density information derived from laboratory experiments was used. This result is not in contradiction with other similar studies such as Kuhn (2003), but suggests that the effect of variable density models on geoid determination may apply only to extensive areas of high terrain and has no significant bearing over small mountainous areas.

Overall, significant progress has been made towards the computation of a high precision gravity model and gravimetric geoid for the Azores area, and towards the unification of the 9 local vertical datums currently in use. This geoid model (AZGEOD07) will enhance future studies of regional ocean circulation and of Earth's dynamics (such as isostatic compensation, admittance computation, or dynamic mantle processes).

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Table Captions

Table 1. Statistics for the gravity anomalies, in mGal (10^{-5}m/s^2). FA is the measured free-air gravity anomalies; Cp is the terrain correction; and RTM are the residual terrain model correction. GGM02C was augmented with EGM96 for degrees 201-360.

Table 2. Statistics for the AZGEOD07 geoid model comparison with mean sea surface minus mean dynamic topography (MSSSH-MDT)-N and with GPS/levelling data over four islands. Local vertical datum offsets are given by the mean of the residuals. Values are in m.

Table 3. Statistics for the comparison of three geoid solutions with GPS/levelling data over four islands. The first solution was computed with a 500 m resolution DTM and standard (2670 kg m^{-3}) density, the second with a 100 m DTM and standard density, and the third with a 100 m DTM and a density of 2423 kg m^{-3} . Values are in m.

Table 4. Statistics for the comparison of the AZGEOD07 geoid model adjusted with a four-parameter model to GPS/levelling data over four islands. There is one four-parameter model for each island. Values are in m.

Tables

Table 1

Gravity anomalies (mGal)	Mean	Std	Min	Max
FA (Free-air anomalies)	36.73	23.34	-82.05	271.96
FA - EGM96 – RTM	-0.16	9.42	-68.60	65.27
FA - GGM02C – RTM	-0.38	9.46	-69.97	64.19
FA – EIGEN_GL04C - RTM	-0.44	10.38	-68.89	69.42
FA - GGM02C + C _p	-0.21	19.45	-104.50	269.44

Table 2

	Mean	Std	Minimum	Maximum
(MSSH-MDT) – N	-0.917	0.138	-1.253	-0.364
(MSSH-MDT) – N (bias + tilt removed)	0.000	0.137	-0.346	0.545
S. Miguel	0.655	0.173	0.349	1.027
Terceira	0.262	0.098	0.016	0.486
Pico	0.120	0.183	-0.281	0.475
Faial	0.147	0.042	0.058	0.222

Table 3

Standard deviation	DTM 500m $\rho = 2670 \text{ kg m}^{-3}$	DTM 100m $\rho = 2670 \text{ kg m}^{-3}$	DTM 100m $\rho = 2423 \text{ kg m}^{-3}$
S. Miguel (77 vertices)	0.207	0.173	0.173
Terceira (59 vertices)	0.113	0.098	0.098
Pico (55 vertices)	0.200	0.183	0.183
Faial (32 vertices)	0.047	0.042	0.042

Table 4

Locally adjusted (Bias + tilt)	Std	Minimum	Maximum
S. Miguel (77 vertices)	0.120	-0.247	0.204
Terceira (59 vertices)	0.095	-0.200	0.350
Pico (55 vertices)	0.120	-0.276	0.261
Faial (32 vertices)	0.021	-0.032	0.047

Figure Captions

1. Gravity data distribution across the Azores archipelago. MAR – Mid-Atlantic Ridge, NAM - North-American plate, EUR - Eurasian plate, AFR – African plate, EAFZ – East Azores Fault Zone, GF – Gloria Fault.
2. Digital terrain model and bathymetric model with a resolution of 100 m.
3. Final free-air gravity anomalies in the area under study.
4. Gravimetric geoid model for the Azores area (AZGEOD07).

Figures

Figure 1

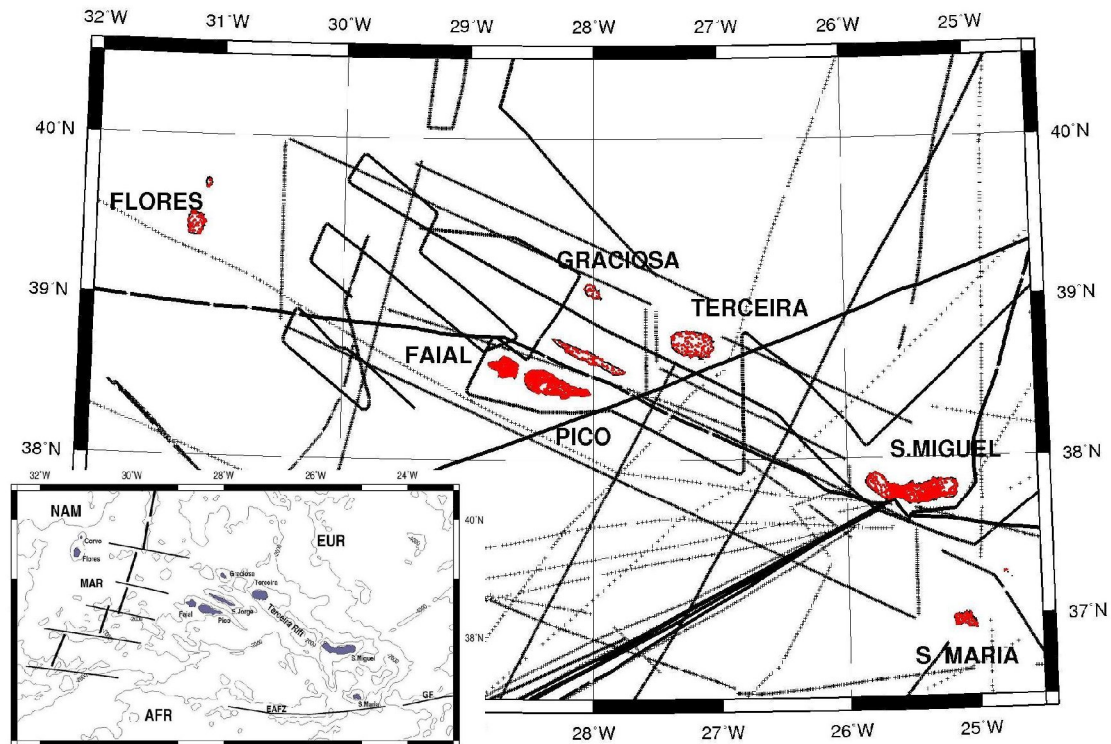


Figure 2

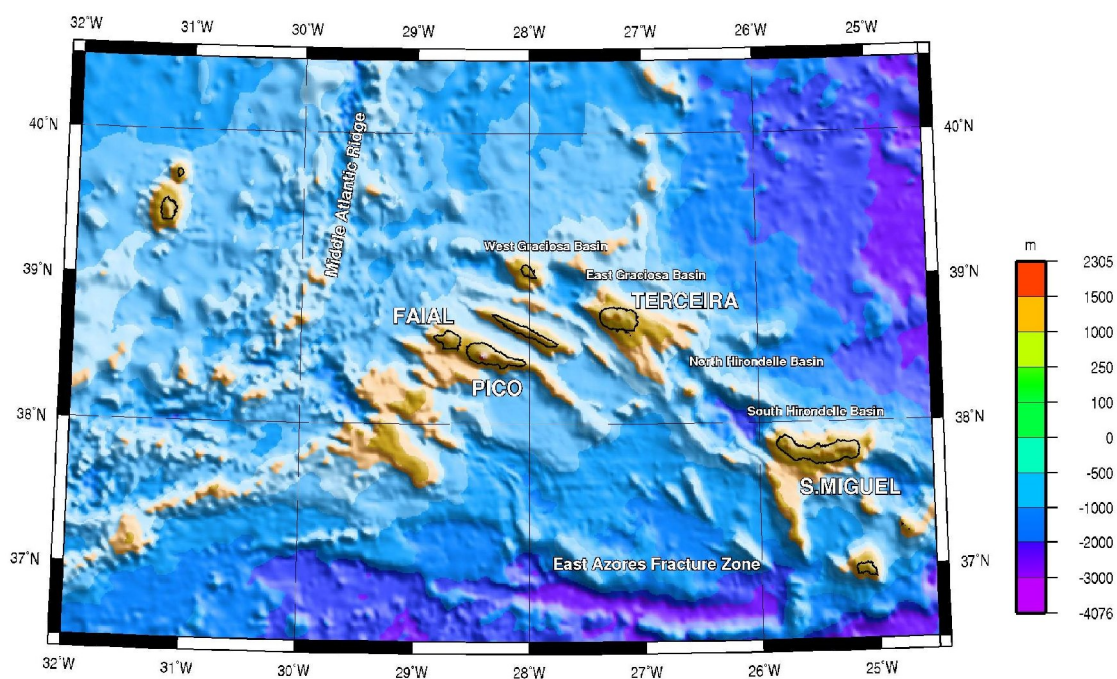


Figure 3

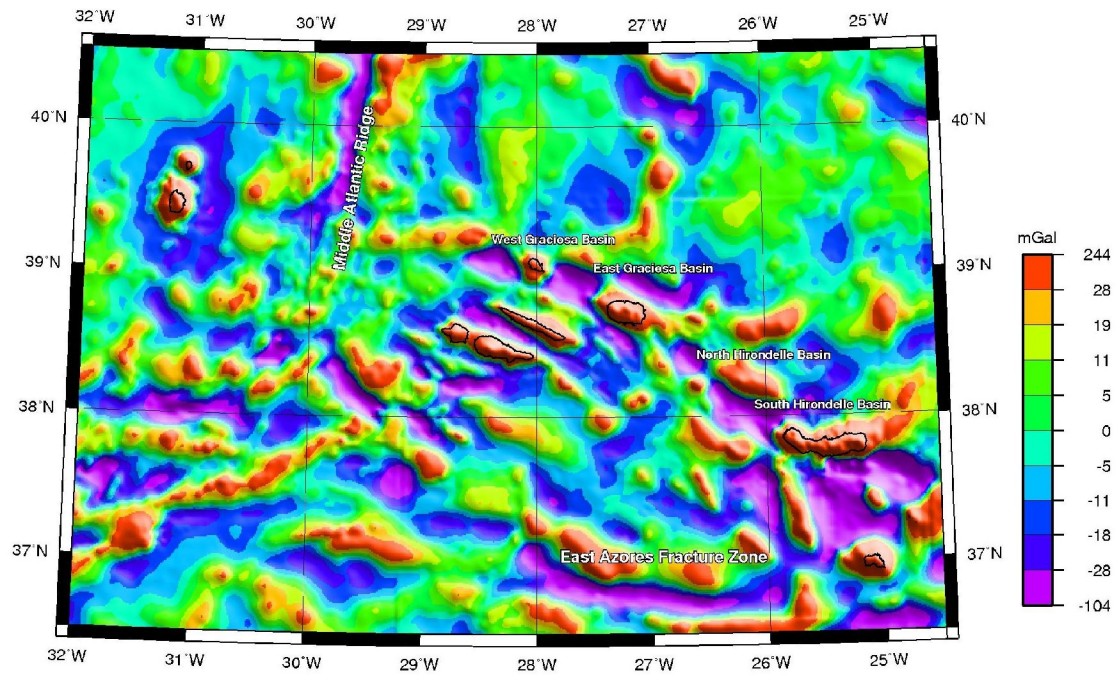


Figure 4

