



Evolution of geoids in recent years and its impact on oceanography

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Summary: Mean surface geostrophic ocean currents may be calculated from the Mean Dynamic Topography (*MDT*), estimated as the difference between a mean sea surface height (*MSS*) calculated from radar altimeters and a reference geoid height. A review of the most widely used geoids is presented. The difference between the third release of the Gravity field and steady-state Ocean Circulation Explorer (*GOCE*) geoid and three earlier geoids (the Earth Geopotential Model 1996 [EGM96], one of the geoids obtained by the Gravity Recovery and Climate Experiment [GRACE05], and the Earth Gravitational Model 2008 [EGM2008]) is computed and interpreted as an 'artefact' *MDT*, i.e. a misfit when non-accurate geoid models are used to calculate the ocean *MDT* and related geostrophic currents. These results are contrasted with the *MDT* computed by comparing the *GOCE* geoid with the *MSS* distributed by Collecte Localisation Satellites in 2001 (CLS01). The comparison shows that there was a strong influence of altimetry measurements in the construction of the EGM96 geoid, i.e. the artefact *MDT* calculated using EGM96 shows a high resemblance to the *MDT* computed using the *MSS* CLS01 field, both considering *GOCE* as the reference geoid. The correlation disappears largely, but not completely, for the two most recent geoids; in particular, the *MSS* has greater global influence on GRACE05 than on EGM2008 although the latter does better at latitudes of less than 60° and is more useful for reproducing the intense western boundary currents. The results show that EGM96 may lead to significant errors in the spatial gradients of *MDT* (for latitudes of less than 60° the global root mean square is 0.2422 m) and therefore in the geostrophic surface velocities. When the spatially averaged GRACE and EGM2008 geoids are used for latitudes of less than 60°, the global *MDT* root mean square is substantially reduced.

Keywords: geoid; mean sea surface; mean dynamic topography; altimetry; surface geostrophic velocity.

Sobre la evolución de los geoides en los últimos años y su impacto en la oceanografía

Resumen: Las corrientes geostróficas superficiales se pueden obtener a partir de la Topografía Dinámica Media (*MDT*), a su vez estimada comparando la altura Media de la Superficie del Mar (*MSS*), medida por altimetría de radar, con la altura del geode de referencia. En este estudio se presenta una reseña de los geoides más usados. Se calcula una *TDM* ficticia a partir de la diferencia entre la tercera versión del geode medido por la misión Gravity field and steady-state Ocean Circulation Explorer (*GOCE*) y tres geoides precedentes: el Earth Geopotential Model 1996 (EGM96), uno de los geoides obtenidos por la misión Gravity Recovery and Climate Experiment (GRACE05) y el Earth Gravitational Model 2008 (EGM2008). Estos resultados se contrastan con la *TDM* calculada comparando el geode de *GOCE* con la *MSS* distribuida por el Collecte Localisation Satellites en 2001 (CLS01). La comparación muestra una fuerte influencia de medidas altimétricas en la síntesis del geode EGM96, i.e. la *MDT* ficticia calculada con el EGM96 es muy parecida a la *MDT* calculada mediante la *MSS* CLS01, usando en ambos casos el geode de *GOCE* como referencia. La correlación desaparece en gran medida, pero no por completo, con los dos geoides más recientes: EGM2008 and GRACE05; en particular, la *MSS* tiene mayor influencia global sobre GRACE05 que sobre EGM2008, aunque este último se comporta mejor para latitudes inferiores a 60°, siendo más adecuado para reproducir las intensas corrientes de frontera oeste. Los resultados muestran que la utilización de EGM96 puede ocasionar errores importantes en los gradientes espaciales de *MDT* (para latitudes inferiores a 60° la media cuadrática global es de 0,2422 m) y, consecuentemente, en las velocidades superficiales geostróficas. Cuando se utilizan los valores promediados espacialmente de GRACE y EGM2008 para latitudes inferiores a 60°, la media cuadrática global de la *MDT* se reduce substancialmente.

Palabras clave: geode; nivel medio del mar; topografía media dinámica; altimetría; velocidad geostrófica superficial.

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INTRODUCTION

A proper determination of the geoid is very important for several disciplines related to both land and sea surface, including geodesy, solid-Earth physics and the subject of this article: oceanography. The geoid is a level surface defined as being everywhere perpendicular to gravity. Therefore, in a motionless ocean the sea surface would be everywhere parallel to the geoid. The shape of the geoid is actually relatively close to the shape of an ellipsoid with an equatorial radius 21.4 km longer than the polar radius (Hughes and Bingham 2008). This ellipsoid gives the plane normal to the local effective gravity, which is no more than the vector addition of the Earth's gravity acceleration and the Earth's centripetal acceleration (a function of latitude, associated with the rotation of the Earth around its axis). However, the geoid may locally depart from this ellipsoid by up to 100 m because of regional changes in the gravitational field (Hughes and Bingham 2008).

The sea surface height (*SSH*), or elevation of the sea surface, is measured globally through satellite altimetry in combination with precise satellite location data. The difference in elevation between *SSH* and the geoid is named the sea surface dynamic topography (*DT*). A moving water-parcel experiences the Coriolis force which, if unbalanced, will drive water displacements that create horizontal changes in the *DT*. This will continue until the Coriolis force is eventually counteracted by the pressure gradient associated with the horizontal variation in *DT*. Alternatively, consider a *DT* perturbation that is created by a transitory process such as sea surface winds or buoyancy fluxes; once this additional force disappears, the associated pressure gradient will accelerate the fluid until the associated Coriolis force is capable of balancing the pressure gradient. In either case, the resulting steady current is said to be in geostrophic balance.

Both the geoid and the *SSH* vary with time but the amplitude as well as the temporal scale of both variations are quite different. The time scale for geoid changes is very long, related to motions in the Earth's lithosphere, so that at the time scales of interest for present day circulation patterns the geoid may be taken as constant. We may speak of total *SSH* as arising from the contributions of mean and anomaly *SSH* values. The mean *SSH* values (or mean sea surface, *MSS*) may be estimated from the average of *SSH* over a relatively long period of time, giving rise to the concept of mean dynamic topography (*MDT*) as the difference of *MSS* less the geoid (*MDT* would be this height difference times the gravity constant but, for simplicity, hereafter we will always refer to it as the height difference, with units of distance). The *MSS* becomes constant in time (over the averaging period) but remains a function of the position over the sea surface, so the *MDT* also changes as a function of position over the sea surface. The difference between total and mean *SSH* gives the

anomaly *SSH*, usually named the sea level anomaly (*SLA*), which is a function of time.

The sea surface *DT* allows us to estimate the surface geostrophic current, or the portion of the surface current in geostrophic balance. The *MDT* tells us about the predominant (quasi-permanent) mean surface geostrophic currents and the *SLA* provides information on the temporal changes, relative to the averaging period, of the geostrophic circulation patterns. This surface geostrophic flow is a principal contribution to the large-scale ocean flow and its near-surface field. This is particularly true for the intense western boundary currents, where the geostrophic contribution in the upper ocean provides for most of the total current. The geostrophic flow is, to a high degree, responsible for redistributing all key climatic properties, including heat and freshwater, therefore being a forcer as well as a tracker of the global climate (e.g. Gill 1982).

The accuracy of the *MDT* depends on the accuracy of both the *MSS* and the geoid. In contrast, the accuracy of the *SLA* does not depend on the determination of the geoid. Consider first the accuracy of the *MSS*. Several studies have shown that the accuracy and spatial resolution of the *MSS* depend on the number of altimetry satellites and the averaging period of the altimetry signal (Hernandez and Schaeffer 2001, Hwang et al. 2002, Rio and Hernandez 2004, Bingham et al. 2008, Andersen and Knudsen 2008, 2010). The definition of *MSS* is certainly not unique, as it depends on the length of the available time series. Because of the relatively recent appearance of satellite altimetry (the first altimeters were SEASAT, launched in June 1978 and operating during 105 days, and GEOSAT, which acquired three years of altimetry data starting in November 1986), the *SSH* averaging period is still relatively short and poses an important limitation on our calculation of the *MSS*. During the last decade a widely used *MSS* has been the one provided by Collecte Localisation Satellites – Centre National d'Etudes Spatiales (CLS-CNES) (Hernandez and Schaeffer 2001). This *MSS* field, hereafter CLS01, was calculated using two-year of data from GEOSAT, five-year data from ERS-1/2 (including all the data acquired by ERS-1 during the geodetic phase), and seven-year data from TOPEX/Poseidon. Hernandez and Schaeffer (2001) conclude that the accuracy of the signal (standard deviation between the model and data fields) is about 1 cm. A more recent study by Andersen and Knudsen (2009) uses a total of 31 years of data from eight different satellites. A comparison of the CLS01 and Andersen and Knudsen (2010) *MSS* fields with independent satellite data (Jason-2 and Envisat) shows that they have similar features, the absolute misfit (*MSS* difference) depending on the satellite used for the comparison. The standard deviation in the *MSS* difference ranges between 1 and 3 cm while the standard deviation in the slope of the *MSS* difference is between 1 and 2 cm km⁻¹ (Schaeffer et

al. 2011). Additional comparisons between CLS01 and the more recent CNES-CLS11 *MSS* fields (Schaeffer et al. 2012) also show differences of the order of 1 cm (Schaeffer et al. 2011).

The second error source in the determination of *MDT* is the geoid itself, particularly at scales of the order of 100 km or less (for the case of GOCE, and even at larger scales for earlier geoids). Several geoid (Earth gravity) models have been used to calculate the *MDT* and the associated mean geostrophic current field (e.g. Hwang et al. 2002, Rio and Hernandez 2004, Andersen and Knudsen 2008, Bingham et al. 2008, Maximenko et al. 2009, Bingham et al. 2011, and Rio et al. 2012). These models have been developed using altimetry and/or gravity field data (see next Section). The determination of the geoid has improved with time, the latest achievement being the Gravity field and steady-state Ocean Circulation Explorer (GOCE) mission (ESA 1999) with a nominal vertical accuracy of 2 cm and a horizontal resolution of 100 km. Since its launch on 17 March 2009, this European Space Agency (ESA) satellite has been continuously acquiring measurements of the Earth's gravity field with unprecedented accuracy and precision (<http://earth.esa.int/GOCE>). The geoid synthesized with this satellite data has been distributed freely to the scientific community since June 2010 (EGG-C 2012) and initial estimates of the *MDT* for the North Atlantic have been provided (Bingham et al. 2011). At the time of this study, three releases of the gravity field have become available, the last one using a total of 16 months of data collected over 18 months (November 2009 - April 2011).

The main objective of our study was to assess how the determination of the geoid has evolved over time and the effect that it has had on the accuracy of the inferred *MDT* and mean geostrophic surface currents fields. The underlying premise is that early geoids constructed using altimetry data were likely contaminated by the *MDT*; even for more recent geoids, which have been obtained using only remotely sensed direct measurements of the Earth gravity field, there may be other error sources. These errors can have significant consequences on the estimation of the *MDT*. To illustrate this, let us name the true geoid *tG*; then the true *MDT* would be obtained as

$$tMDT = tMSS - tG \quad (1)$$

where *tMSS* is the true *MSS*, which is calculated after a certain period of altimetry measurements, here considered to be constant in time.

In reality, however, we do not have the true geoid but an approximate one, *G*, so we can only calculate an approximate *MDT* as

$$MDT = tMSS - G \quad (2)$$

Rearranging and subtracting *tG* on both sides of this expression leads to

$$MDT + (G - tG) = tMSS - tG \quad (3)$$

Therefore, because of Eq. 1, we have

$$MDT + aMDT = tMDT \quad (4)$$

where we have defined $aMDT = G - tG$ as the difference between the measured and true geoids. Equation 4 tells us that the *tMDT* is equal to the measured *MDT* plus a quantity *aMDT* which we identify as an artefact *MDT*. In the above discussion we have ignored inaccuracies in *MSS* arising from instrumental errors and the limited time extent of altimetry measurements. The *MSS* estimates have improved in time mainly thanks to the availability of more altimetry data, e.g. from Jason-1, Jason-2, and Envisat. We will not deal with them, but will simply consider our best available estimate for *MSS* and examine the error that is associated with the inaccuracies in the geoid field.

We do not have anything like a true geoid but yet we may assume that their estimation has progressively improved in time. Therefore, we will assume the third (and so far last) release of the GOCE geoid to be the true geoid. As shown above, the differences between the earlier—and possibly less accurate—geoids and GOCE can have significant consequences on the estimation of *MDT*. Specifically, we will examine three widely employed early geoids, namely the Earth Geopotential Model 1996 (EGM96) geoid (Lemoine et al. 1998), one of the geoids obtained by the Gravity Recovery and Climate Experiment (GRACE05) (Tapley et al. 2004), and the Earth Gravitational Model 2008 (EGM2008) (Pavlis et al. 2008). The differences between any of these early geoids and the last release of GOCE will give the *aMDT* (Eq. 4). The underlying hypothesis is that *aMDT* should decrease in time, as the most recent Earth gravity models have been calculated without relying on altimetry measurements, and asymptotically tend to the “true” geoid. Any significant differences between recent models may be indicative of remaining inaccuracies in the computed *MDT*.

The *aMDT* will be compared with our best-estimate *MDT* as obtained using GOCE together with the CLS01 *MSS* data (Hernandez and Schaeffer 2001). As these *MSS* data are independent from the GOCE data, respectively obtained from altimetry and gravity measurements, we may expect that the variations in *MDT* computed from the CLS01 *MSS* and GOCE fields should respond to dynamic processes and bear little or no correlation with the *aMDT*. Possible signals in *aMDT* and correlations with the best-estimate *MSS* will be discussed.

The rest of the paper is structured as follows. The geoids as well as the *MSS* field are first described and the methodology followed to compute the residual *MDT* and surface geostrophic velocities is briefly explained. The main results, as obtained from the comparison between early geoids and GOCE, are then presented. The paper ends with some concluding remarks.

GEOID AND ALTIMETRY DATA

Four different geoids are used in this study:

- EGM96: The Earth Geopotential Model 1996 embodies ground-based as well as satellite measurements (Lemoine et al. 1998). Remotely sensed data were mainly radar altimeter observations, averaged

and transformed in gravity anomalies as expressed in Sandwell and Smith (1997). This geoid may be downloaded from the National Aeronautics and Space Administration (NASA) Goddard Space Flight Center web page (<http://cdis.gsfc.nasa.gov/926/egm96/egm96.html>). In this study a 360-degree and -order version of the EGM96, included in the GOCE User Toolbox (GUT), was used. GUT is distributed by the ESA at <http://earth.esa.int/gut/>. In the spatial domain, the geoid is defined on a 0.5° uniform grid. Since the calculation of the EGM96 geoid included altimetry measurements, we may expect that it might have been influenced by some signal from the *MSS* fields.

- GRACE05: The Gravity Recovery and Climate Experiment (GRACE) (Tapley et al. 2004) is a joint mission by the American (NASA) and the German (DLR) space agencies, launched in March 2002 with the objective of making detailed measurements of the Earth's gravity field. GRACE's data are available on the web pages of these two institutions (<http://podaac.jpl.nasa.gov/grace> and <http://isdc.gfz-potsdam.de/grace>). As for EGM96, the GRACE05 geoid used in this study is included in the GUT (see Foerste et al. [2008]). It is characterized by a spatial resolution of 0.5° and a spherical harmonic degree and order 360. GRACE05, in the version used for this study, includes altimetry measurements.

- EGM2008: The Earth Gravitational Model 2008 is described in Pavlis et al. (2008) and distributed by the International Centre for Global Gravity Field Models (<http://icgem.gfz-potsdam.de/ICGEM/shms/egm2008.gfc>). It is available up to spherical harmonic degree and order 2159, and contains additional coefficients extending to degree 2190 and order 2159. For consistency with the EGM96 and GRACE05, the geoid heights have been calculated using up to the spherical harmonic degree and order 360, with a spatial resolution of 0.5° . Like EGM96, EGM2008 is the outcome of fusing different data from several sources, including altimetry data.

- GOCE's level 2 time-wise product: The time-wise product is obtained using only direct measurements of the gravity field, without any altimetry contribution. Details about the GOCE mission can be found on the ESA web page (<http://earth.esa.int/GOCE>) and in the ESA (1999) report; data may be accessed on the ESA web page. Three different products are available, namely the direct solution, the space-wise solution and the time-wise solution; the last of these was used in this study as it does not rely on auxiliary information (EGG-C 2012). We used the third release, the last one available at the time of the study, based on 16 months of data collected over 18 months (from November 2009 to April 2011). The spatial resolution of the GOCE level 2 products is 0.5° , with degree and order 250.

Additionally, as mentioned in the introduction, the tidal-free *MSS* field is determined from the CLS01 algorithms (Hernandez and Schaeffer 2001). A two-minute version of this field, included in the GUT, is used to derive a 250-degree and -order version of the *MSS* with 0.5° spatial resolution.

All four geoids, as well as the *SSH* field, are referred to the same geodetic reference system, the GRS80 el-

lipoid, consisting of a global reference ellipsoid that sets a model of the Earth's gravity field (Moritz 2000).

METHODOLOGY

To assess the evolution of the geoid over the years, the EGM96, GRACE05 and EGM2008 geoids are compared with the latest release of the GOCE geoid. The differences may be caused by several factors, including different input measurements (with or without altimetry data), changing temporal windows and differences in signal processing techniques. Regardless of the cause of misfit, when used for *MDT* calculations, the differences between geoids may be considered as an *aMDT*, an artefact value that can be calculated as follows

$$\eta_{art} = h_{mod} - h_{GOCE} \quad (5)$$

where h_{mod} and h_{GOCE} are the heights from the former Earth gravity models (EGM96, GRACE05 and EGM2008) and from the GOCE Earth gravity model, respectively. This *aMDT* may be compared with our best estimate for the *MDT*, calculated as

$$\eta_{ref} = h_{CLS} - h_{GOCE}, \quad (6)$$

where h_{CLS} is the sea surface height as obtained from the CLS01 *MSS* field.

The zonal and meridional components of the geostrophic surface currents (u_g, v_g) are calculated from either the artefact ($\eta = \eta_{art}$) or best-estimate ($\eta = \eta_{ref}$) *MDT* as follows

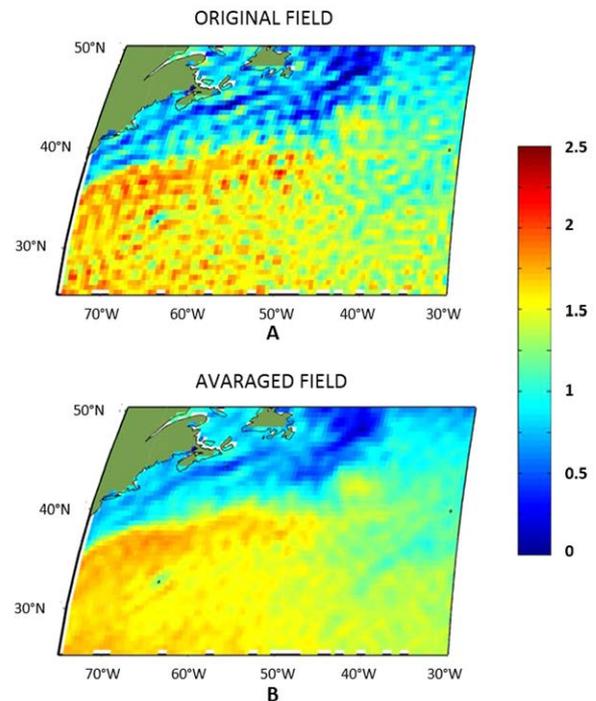


Fig. 1. – Best-estimate *MDT* [m] obtained with the CLS01 *MSS* and GOCE geoid for the NWA region. (a) before and (b) after the spatial averaging.

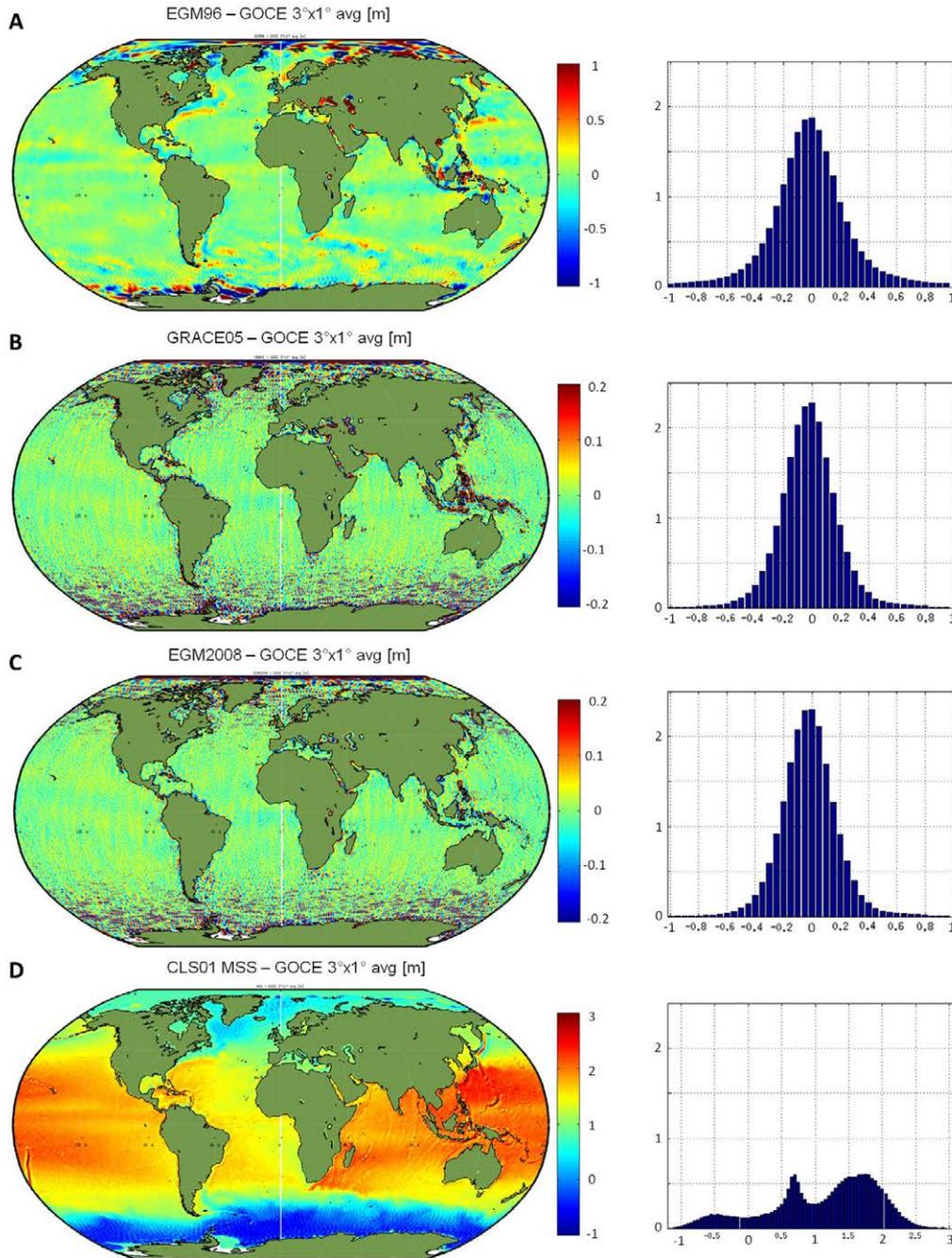


Fig. 2. – Global maps and histograms for the averaged values of three *aMDT* (comparing the EGM96, GRACE05 and EGM2008 models to the GOCE model) and the best-estimate *MDT* (comparing the CLS01 *MSS* field with the GOCE model); values are presented in [m]. Note the changing colour scale in the different panels.

$$(u_g, v_g) = \frac{g}{f} \left(-\frac{\partial \eta}{\partial y}, \frac{\partial \eta}{\partial x} \right) \quad (7)$$

where x and y are the horizontal distances along parallels and meridians, g is the gravitational acceleration, $g = 9.780327(1+0.0053024 \sin^2 \theta - 0.0000058 \sin^2 2\theta)$ m s⁻², with θ the latitude, and f is the Coriolis parameter, $f = 2\Omega \sin \theta$, with Ω the angular velocity of the Earth. The magnitude of the geostrophic velocity field is computed as $(u_g^2 + v_g^2)^{1/2}$.

ARTEFACT AND BEST-ESTIMATE MEAN DYNAMIC TOPOGRAPHIES

Mean dynamic topography

We may now proceed to calculate the artefact and best-estimate *MDTs*, the former using the older geoids as compared with GOCE and the latter with the CLS01 *MSS* field and GOCE’s geoid (Eq. 6). Due to the definition of the geoid as a finite series of spherical harmonics, small undulatory modulations of the *MDT* are ob-

served all over the globe (Bingham et al. 2008). Since our objective is to highlight the differences between the geoids, and how these can affect the geostrophic current calculation, our first step is to remove these undulations. With this purpose we have followed a simple spatial averaging: each point is substituted by the mean value inside a rectangular window of 3×1 degrees (in longitude and latitude, respectively) centred on the grid point itself; such a rectangular window takes into account the greater zonal than meridional coherence of mean surface ocean currents.

Figure 1 illustrates the effect of spatial averaging on the northwest Atlantic (NWA), bound in latitude and longitude by (75°W , 30°W) and (25°N , 50°N). Averaging adequately removes the short-scale undulations at the expense of moderately reducing the absolute maximum values. The Gulf Stream signal, as it is characterized by a spatial scale substantially larger than the small scale anomalies (approximately 10° or ten times larger than the anomalies), remains clear.

Figure 2 presents global maps, as well as histograms, for both the artefact and the best-estimate averaged *MDTs*. The histograms show the existence of non-zero artefact *MDTs*, with the absolute values decreasing from EGM96 to EGM2008 and to GRACE05. In all cases the largest differences are related to high latitudes and the presence of strong western boundary currents, such as the Gulf Stream. The large differences in the EGM96 and GOCE geoids are possibly due to the inclusion of *MSS* information in the definition of the EGM96 geoid. The best-estimate *MDT* has a rather irregular latitudinal distribution, with a mean positive bias due to the much greater surface area of the warm tropical ocean. These results confirm a progressive general improvement of the geoid, with a substantial change from EGM96 to EGM2008 and much less between EGM2008 and GRACE05; EGM2008 is, in fact, based on GRACE at these scales (3×1 degrees).

As an additional comparison, the root mean square (rms) of the artefact and best-estimate averaged *MDTs* were calculated and are plotted in Figure 3 as a function of the latitude. As can be seen, the best-estimate *MDT* behaves differently from the artefact *MDT*, reaching its maximum rms values at the tropical and equatorial regions. On the other hand, all *aMDT* have a similar pattern, with maximum rms values in the polar regions and a quasi-constant shape for latitudes of less than 60° , being noisier in the southern hemisphere. The high artefact values in the polar regions may be due to less availability of altimetry data as a result of sea ice coverage but they also probably reflect the appearance of short-scale variability, which is not well removed by the 3° by 1° averaging process; i.e. at 60°N the distance for 1° of latitude is only about 50 km, less than the resolution of the EGM2008 and GRACE05 at degree/order 360 (55 km) and GOCE at degree/order 250 (80 km). Concerning the absolute values, mean rms values are summarized in Table 1. Two sets of values are shown: the first uses all grid points in the global ocean and the second uses only those grid points at latitudes of less than 60° ; a significant improvement in the quality of the results is observed when the high-latitude values

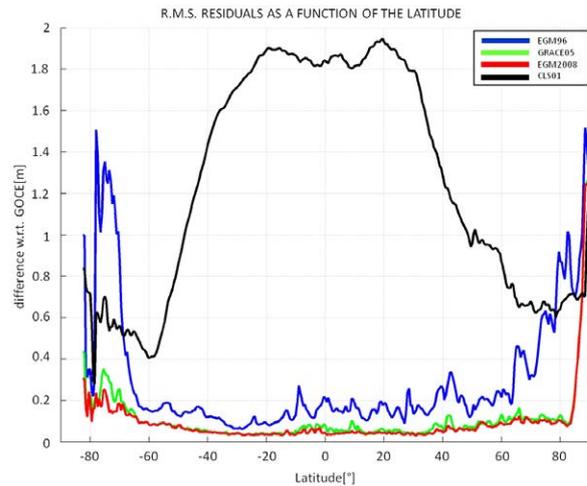


Fig. 3. – Latitudinal dependence of the zonally-averaged artefact and best-estimate *MDT* rms values.

are removed. According to Table 1, and as shown in Figure 2, GRACE05 and EGM2008 have very similar performances, with EGM2008 (0.1859 m) just a little smaller than GRACE05 (0.1932 m), both values being less than the EGM96 value (0.3990 m). When only latitudes of less than 60° are considered, the rms decreases substantially for all geoids, with EGM2008 (0.0536 m) slightly less than GRACE05 (0.0625 m) and yet substantially smaller than EGM96 (0.1553 m).

Table 1 also presents the statistics for the best-estimate *MDT*. As expected, the corresponding rms is much larger than the *aMDT*, with a difference of one order of magnitude either considering all the points or only data for latitudes less than 60° (2.080 and 2.698 m, respectively), reflecting the existence of real physical processes.

Geostrophic currents

The intensity of the artefact currents, as obtained using the averaged *aMDTs*, may be contrasted with their intensity as determined using the best-estimate averaged *MDT* (Fig. 4); a zonal band of 10° of latitude around the equator has been blanked: in this region the Coriolis parameter is very small, so very large velocities would be required in order for the Coriolis force to be significant. In other words, there are other forces larger than the Coriolis force, so the dynamics is far from geostrophic.

When the EGM96 model is used, there are significant *aMDT* gradients that stand out from the general noisy background field (Fig. 4A), which are

Table 1. – Root mean square (rms) values for the artefact *MDT* using the EGM96, GRACE, and EGM2008 gravity models and for the best-estimate *MDT* using CLS01 and GOCE. Bold figures are obtained using only points at latitudes less than 60° .

	rms [m]
EGM96 – GOCE	0.399 / 0.155
GRACE05 – GOCE	0.193 / 0.063
EGM2008 – GOCE	0.186 / 0.054
CLS01 <i>MSS</i> – GOCE	2.080 / 2.698

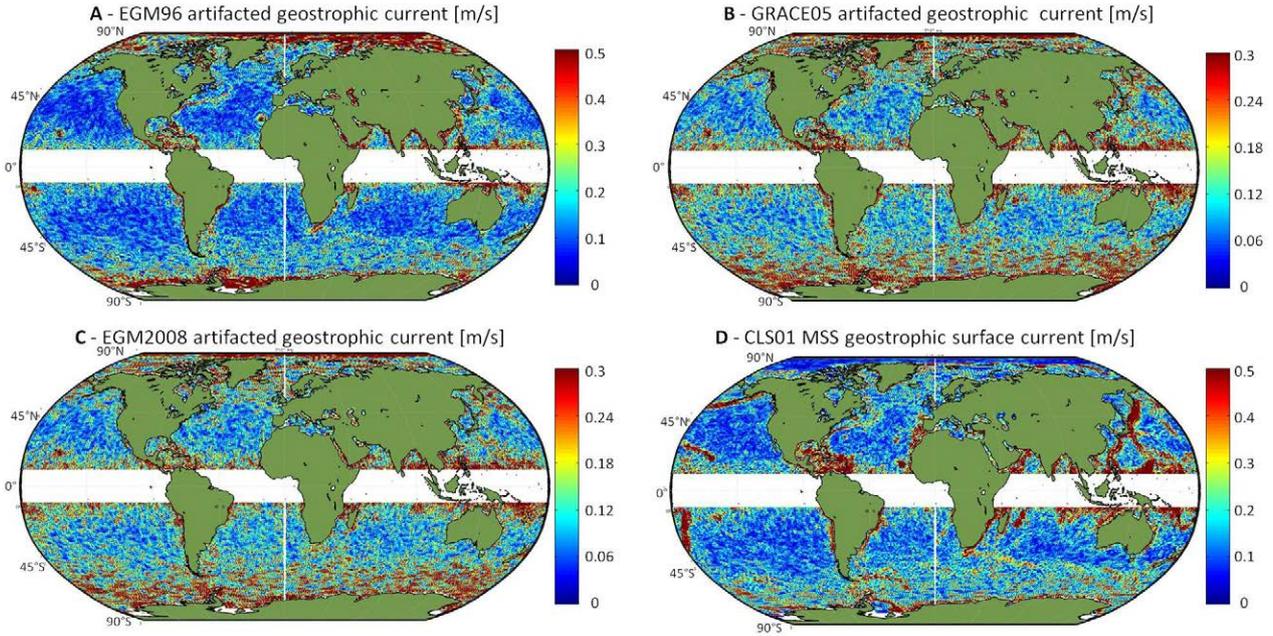


Fig. 4. – Intensity of the surface geostrophic currents: artefact currents calculated using the (a) EGM96, (b) GRACE, (c) EGM2008 geoids, and best-estimate currents calculated using (d) the CLS01 *MSS* values. In all cases the values are expressed in [m/s].

in correspondence with some of the major oceanic currents, such as the western boundary currents and the Agulhas Current (Fig. 4D); this strongly suggests that the EGM96 geoid was contaminated by the assimilation of altimetry data. In contrast, the artefact fields obtained using the GRACE05 and EGM2008 geoids are very similar to each other, lacking any noticeable correspondence with the *MDT* gradients associated with the western boundary currents (Fig. 4B,C), at least at low-mid latitudes and the spatial scales of this study.

ANALYSIS FOR THE NORTHWEST ATLANTIC REGION

The above results indicate that there has been a substantial improvement in the definition of the geoid over time, particularly from the early EGM96 to the GRACE05 and EGM2008 models, but yet they raise some questions about possible local errors. In an attempt to assess the source and size of these errors, we examined the artefact and best-estimate raw and averaged *MDT* for the selected area in the NWA (Fig. 1). A

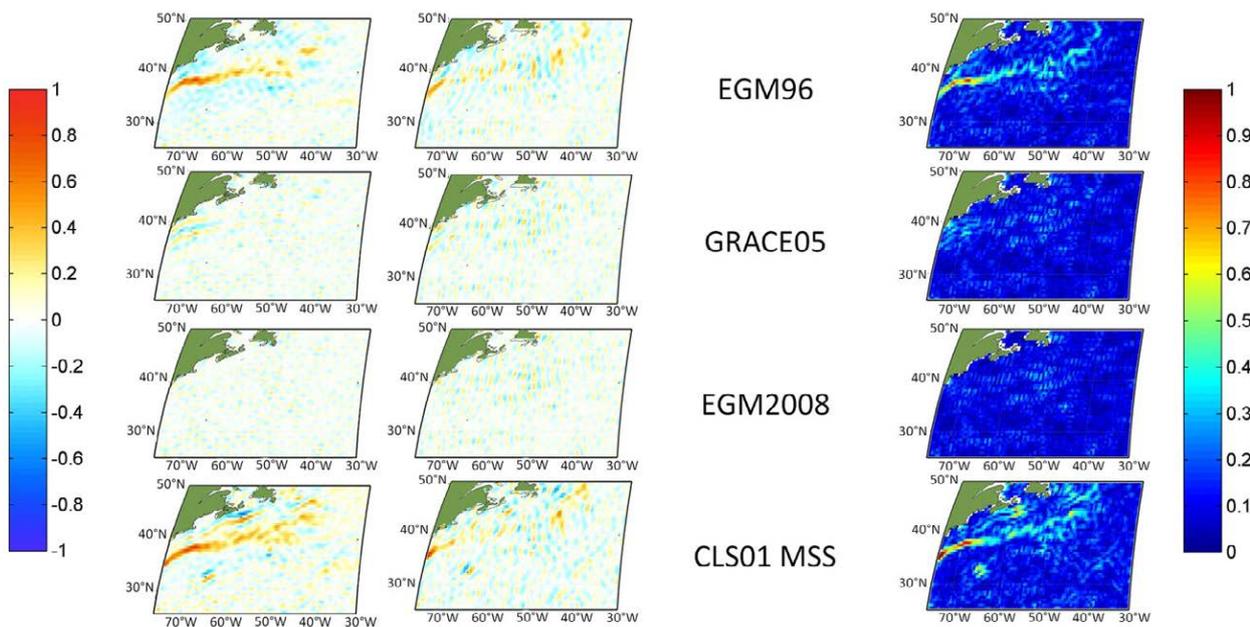


Fig. 5. – Zoom to the zonal component, meridional component and total surface geostrophic currents in the NWA region: artefact currents calculated using the (a) EGM96, (b) GRACE, (c) EGM2008 geoids, and best-estimate currents calculated using (d) the CLS01 *MSS* field. In all cases the values are expressed in [m/s].

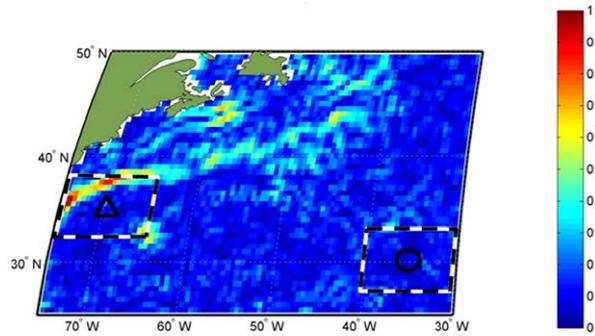


Fig. 6. – Data subsets within the NWA region: The triangle area is affected by the Gulf Stream while the circle area is a stable zone near the centre of the subtropical gyre. The background map corresponds to the best-estimate mean currents as calculated using the CLS01 MSS field (m s^{-1}).

detailed view of the zonal and meridional components as obtained from the averaged *MDT* values, as well as the total intensity, is presented in Figure 5. The EGM96 geoid leads to the highest velocities, with strong zonal velocity components in the Gulf Stream area that lead to maximum speeds above 0.5 m s^{-1} . The artefact Gulf Stream currents are substantially reduced for the GRACE05 residual (no more than $0.1\text{-}0.2 \text{ m s}^{-1}$), and almost disappear with the EGM2008 residual.

Figure 5 shows that within the NWA there are two quite different dynamical regimes: weak currents within the interior gyre of the North Atlantic subtropical gyre and an intense western boundary current, namely the Gulf Stream (e.g. Schmitz and McCartney 1993). Therefore, it is useful to select data subsets from two different areas as follows:

- The first one corresponds to an area strongly affected by the Gulf Stream, between $33^\circ\text{N } 74^\circ\text{W}$ and $38^\circ\text{N } 64^\circ\text{W}$. In this region the spatial *MDT* gradients are potentially large and the correlation between the artefact and best-estimate *MDT* may locally be high (triangle area in Fig. 6). A smaller sub-area, of size one-fourth of the Gulf Stream area and comprising only the swift waters of the Gulf Stream is limited by the points $36^\circ\text{N } 74^\circ\text{W}$ and $38^\circ\text{N } 69^\circ\text{W}$; it corresponds to the upper-left corner of the triangle area in Figure 6.
- The second one is a dynamically tranquil area not affected by the Gulf Stream and located near the centre

of the North Atlantic subtropical gyre, here selected as between $28^\circ\text{N } 40^\circ\text{W}$ and $33^\circ\text{N } 30^\circ\text{W}$. In this area the spatial *MDT* gradients should be small, close to zero, so the dispersion may be indicative of intrinsic limitations in the resolution of the data (circle area in Fig. 6).

The linear relation between the artefact and best-estimate *MDT* indeed changes depending on the selected geoid and area. Figure 7 presents scatter plots for *aMDT* as a function of the best-estimate *MDT*, using averaged data for the whole NWA region. On these plots we have identified those points corresponding to the Gulf Stream and central gyre areas; linear fits (with mean slope η_{art}/η_{ref}) for the whole region and the two separate areas have been included. The linear fit for the Gulf Stream changes dramatically from the oldest to the more recent geoids. For the oldest geoid, EGM96, the data points display a slope of 0.695 (Fig. 7A and Table 2); this slope would increase to 0.847 within the reduced Gulf Stream sub-area, getting closer to the unit slope which would correspond to a geoid identical to the *MSS* (Table 2). The situation changes noticeably with the GRACE05 and EGM2008 geoids, with a progressively decreasing slope (Fig. 7B,C). These results indeed confirm that the EGM96 geoid was highly contaminated by the *MSS* altimetry signal in this intense western boundary current, in contrast to the other two more modern geoids.

The linear fit for the central gyre area behaves in a similar way, with the slope of η_{art}/η_{ref} remaining relatively large for EGM96 (0.403) and decreasing to about 0.1 for the other two geoids (Fig. 7B,C and Table 2). It is perhaps surprising that, for EGM2008, the slope in the central gyre area is significantly larger than in the Gulf Stream area (0.104 as compared with 0.026, Table 2). Nevertheless, this is to be interpreted only as posing a limitation in the height-resolution for an area where the absolute changes in the best-estimate *MDT* are already small (of the order of 2 cm, see below).

The statistics for the NWA, as a whole and separately for both areas, are reported in Tables 2 and 3 both before and after applying the $3 \times 1^\circ$ averaging; in Table 2 we report the ratio η_{art}/η_{ref} while in Table 3 we give the results for the standard deviation of η_{art} with respect to either zero or the linear fit. Consider first the results for the η_{art}/η_{ref} ratio (Table 2). As we

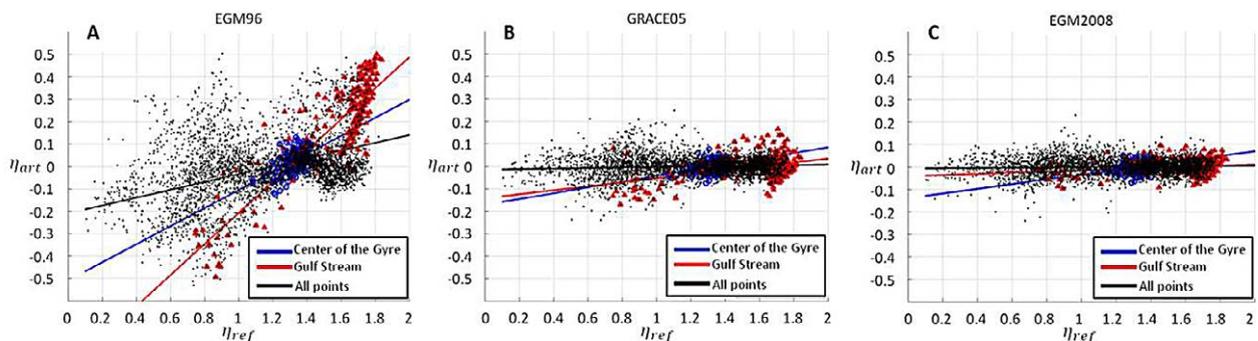


Fig. 7. – Scatter plots of the *MDT* misfit as obtained using the averaged data in the NWA region. (a) EGM96, (b) GRACE05, and (c) EGM2008 geoids, as a function of the best-estimate *MDT* [m]. Red triangles indicate points affected by the Gulf Stream, empty blue and green circles are points in the centre of the subtropical gyre (Fig. 6); the corresponding colour-coded regression lines are superimposed.

Table 2. – Summary of the evolution of the η_{art}/η_{ref} ratio for the three geoids and the CLS01 *MSS*. The Gulf Stream and central gyre areas respectively correspond to triangles and circles in Figure 6; the values for the small Gulf Stream sub-area are shown inside parenthesis.

Values w.r.t. GOCE			EGM96	GRACE	EGM2008	CLS01
NWA region	η_{art}/η_{ref}	Non-aver	0.270	0.127	0.120	1
		Aver	0.175	0.011	0.006	1
Gulf Stream area	η_{art}/η_{ref}	Non-aver	0.747 (0.837)	0.282 (0.290)	0.195 (0.130)	1
		Aver	0.695 (0.847)	0.088 (0.167)	0.026 (0.044)	1
Central gyre area	η_{art}/η_{ref}	Non-aver	0.763	0.686	0.674	1
		Aver	0.403	0.127	0.104	1

Table 3. – Summary of the evolution of the standard deviation of η_{art} for the three geoids and the CLS01 *MSS*. The Gulf Stream and central gyre areas respectively correspond to triangles and circles in Figure 6; the values for the small Gulf Stream sub-area are shown inside parenthesis.

Values w.r.t. GOCE			EGM96	GRACE	EGM2008	CLS01
Gulf Stream area	σ for η_{art} [m]	Non-aver	0.300 (0.433)	0.188 (0.232)	0.170 (0.179)	0.321 (0.437)
		Aver	0.214 (0.322)	0.060 (0.090)	0.037 (0.044)	0.269 (0.362)
	σ [m] w.r.t. linear fit	Non-aver	0.180 (0.233)	0.146 (0.195)	0.143 (0.170)	0 (0)
		Aver	0.104 (0.098)	0.055 (0.067)	0.036 (0.041)	0 (0)
Central gyre area	σ for η_{art} [m]	Non-aver	0.140	0.132	0.133	0.138
		Aver	0.046	0.025	0.023	0.064
	σ [m] w.r.t. linear fit	Non-aver	0.091	0.092	0.094	0
		Aver	0.039	0.024	0.022	0

use the more recent results, the region as a whole experiences a progressive reduction of the artefact slope, which stands out most clearly in the averaged values. Nevertheless, the undulatory character of all geoids shows up clearly in the raw data for the central gyre area, where the variability is of the same order as the relatively small dynamic signal (Fig. 1).

Consider now the standard deviations of η_{art} with respect to either zero or the linear fit for both areas (Table 3). For the Gulf Stream area, there is a substantial decrease of the raw, and particularly the averaged, values as we switch from EGM96 to GRACE05 and to EGM2008; the standard deviation for the EGM96 averaged values is 21 cm while for EGM2008 it has decreased to about 4 cm. For the small Gulf Stream sub-area the relative reduction is even greater, from 32 to 4 cm. In contrast, in the central gyre area, the η_{art} standard deviation changes little as the calculations are carried out using more recent geoids. The non-averaged values are several times larger than the averaged ones, the latter reaching a minimum value of approximately 2 cm for EGM2008. These values remain very close to the accuracy of both *SSH* and gravity models data (between 2 and 3 cm), confirming the stability of the central gyre area and thus validating GOCE's measurements.

Table 3 also provides the standard deviations for the best-estimate raw and averaged *MDT*. As mentioned above, the non-averaged values clearly remain largely affected by the undulations; this happens for all geoids, indicating that this limitation, at least to the 360 degree and order here selected, also affects GRACE05. After averaging, the best-estimate *MDT* numbers are just slightly larger than the *aMDT* values for EGM96 but the difference increases greatly as we move to the two more modern geoids, with EGM2008 performing best. As a reference value, we note that a change in elevation of 6 cm at a latitude of 30°, as observed by CLS01 for the central gyre area, acting over a distance of 963 km (the longest side of the box), would imply surface geostrophic currents of about 1 cm s⁻¹, about what is expected for the centre of the gyre.

CONCLUDING REMARKS

In this study, the most widely used geoids, namely the Earth Geopotential Model 1996 (EGM96), the GRACE05 geoid and the Earth Gravitational Model 2008 (EGM2008), have been compared with the third release (the last one so far) of the Gravity field and steady-state Ocean Circulation Explorer (GOCE) geoid. The comparison has been performed in terms of misfit in estimating the sea-surface mean dynamic topography (*MDT*) and geostrophic currents. The GOCE geoid has been used as a reference and EGM96, GRACE05, and EGM2008 have been compared with it. The differences between the geoids have been translated into artefact geostrophic currents and the main features have been discussed. *MDT* and geostrophic currents have also been calculated using the mean sea surface field distributed by the Collecte Localisation Satellites – Centre National d'Etudes Spatiales (CLS-CNES) minus GOCE geoid model, as an element of comparison.

The raw geoids have been calculated with a resolution of 0.5°×0.5°, with spherical harmonics of 360 degree and order. The results display significant undulations in most oceanic regions which are as large as the EGM96 residuals across the Gulf Stream (about 30 cm); these have been removed by spatial averaging over a moving window of 3° in longitude by 1° in latitude. The resulting averaged fields show a progressive improvement as the artefact *MDT* (*aMDT*) is calculated using the more recent geoid models. EGM96 is clearly affected by the mean sea surface (*MSS*) but this contamination is substantially reduced in GRACE05 and EGM2008. Both GRACE05 and EGM2008 do similarly well: globally GRACE05 behaves slightly better than EGM2008 but the latter does better at latitudes of less than 60°N.

The above results are confirmed by the scatterplot of *aMDT* as compared with our best-estimate *MDT*, showing a very high correlation between the geoid and the CLS01 *MSS* for EGM96, which decreases for GRACE05 and EGM2008. An analysis of Gulf Stream

data quantifies that *aMDT* values associated with EGM96 would lead to errors in excess of 0.1 m s^{-1} , and substantially less for GRACE (0.03 m s^{-1}) and EGM2008 (0.015 m s^{-1}). This EGM2008 value is appraised to be within the noise uncertainty associated with either the *MSS* or *GOCE*, as estimated for quiescent subtropical gyre areas.

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