IMPACT OF GEOID DEFINITION ON THE SIMULATED STRENGTH OF THE EQUATORIAL UNDERCURRENT IN A GLOBAL DATA ASSIMILATION SYSTEM

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ABSTRACT

The availability of gravity data from recent satellite missions (CHAMP, GRACE) and the future GOCE mission will improve our current knowledge of the Earth’s geoid. An accurate description of the geoid is essential for a proper interpretation of altimetric sea level data, which can help us to better understand the ocean circulation. This study evaluates the sensitivity of an estimated ocean circulation to the definition of the geoid, using a global ocean circulation model (OGCM) and an advanced data assimilation technique.

In the analysis system, TOPEX/Poseidon altimetric sea level heights are related to modeled sea level. Instead of using sea level anomalies with respect to a long-term mean, which is the most common usage of altimetry, the absolute sea level minus the geoid estimate is assimilated. In the current study, we adopt TOPEX/Poseidon (T/P) altimetry and incorporate one of the most recent geoid solutions based on satellite data (GGM01C, derived from GRACE observations).

The analysis system uses the OPA OGCM, in which observations are being assimilated. The assimilation technique is an ensemble-based method, that determines the optimal model trajectory by statistical evaluation of a number of perturbed model simulations. The implementation of this technique is especially efficient in a parallel computing environment.

Specific attention is being paid to the equatorial undercurrent. We evaluate the impact of the geoid definition on its strength, and discuss the implications of improving geoid accuracies for the estimation of ocean circulation.

1 INTRODUCTION

Ocean circulation is strongly related to the wind-stress forcing, heat- and fresh water fluxes, which together maintain and modify the transport of water masses across the globe. Although many observations are available of the ocean circulation and its forcing, little is known about the exact magnitude of large scale currents, about the conditions for their existence, and what determines their variability at different time scales.

One of the problems with the interpretation of observations is the lack of a long, continuous and globally homogeneous data set. Although the number of observations has largely increased over the past 50 years, observations are often irregular samples in time and space. Only few sections of in-situ observations of temperature and salinity capture the entire water column, which makes it almost impossible to determine the total transport across a section or the total velocity field. The development of various data assimilation techniques has helped the interpretation of in-situ observations, while the introduction of satellite observations has improved the global coverage. In particular the advent of altimetry has facilitated the monitoring of sea-level variability and associated changes in ocean circulation. However, for a quantitative estimate of ocean transports, an accurate reference level is still missing.

This reference level, or marine geoid height is thus of great importance to the estimations of the mean ocean circulation and the associated water- and heat transports in the global oceans. With the Gravity Field and Steady-State Ocean Circulation Explorer (GOCE) that is planned to be launched in 2006, a precise estimate of the marine geoid height is expected to become available. Combining these gravity data with altimetric sea-level data (e.g. TOPEX/Poseidon, T/P) we will be able to estimate the dynamic topography. The gradients in this topography are a measure for geostrophic (surface) currents, and when used in a data assimilation system, the sea level condition given by the dynamic topography constrains the three-dimensional velocity field. Accuracy in the dynamic topography depends on the accuracy of its two components: the geoid (the gravity observations) and the mean sea level topography (sea level observations).

Recent applications of altimetry in the equatorial Pacific Ocean suggest that the simulation of the equatorial (thermo)dynamics is highly sensitive to the definition of the reference level of the assimilated sea level anomalies [12]. This sensitivity will be studied in the present paper, using the GGM01C geoid derived from GRACE observations in combination with historical tracking data, terrestrial gravity data and satellite altimetry [10] in comparison to the EGM96 geoid [5]. Of particular interest is the simulation of the equatorial undercurrent subject to differences in the marine geoid height.
2 METHODOLOGY

To study the impact of long-time mean sea level variations on the model simulation of ocean currents, we perform data assimilation experiments. In these experiments, mean sea level $\Psi$ is derived from altimetric sea level in combination with a geoid estimate following:

$$\Psi = \langle \eta \rangle - \langle G \rangle,$$

(1)

where $\eta$ denotes T/P altimetric observations, $G$ geoid heights, and the overbar denotes an average over all available data (in this case: 8 years of T/P observations). Note that T/P observations and geoid heights are relative to the same reference ellipsoid.

The model equivalent of this mean sea level, or dynamic topography $\Psi_{\text{mod}}$, is obtained from model output by taking the average over the available time series of model sea level. The remaining residuals $\Psi'$ in the model are the sea level anomalies to be compared to altimetric anomalies $(\eta - \langle \eta \rangle)$, while the model mean sea level $\Psi_{\text{mod}}$ is compared to the observed mean dynamic topography, i.e. $\langle \eta \rangle - \langle G \rangle$. By using two different geoid models for $\langle G \rangle$, we study the impact of the mean sea level on the simulation of ocean dynamics.

As a first step, we choose two different dynamic topographies (i.e. a ‘GGM01C’ dynamic topography and an ‘EGM96’ dynamic topography). Denoting the sea level relative to GGM01C with $\Psi_{GGM}$, and relative to EGM96 with $\Psi_{EGM}$, we can monitor the sensitivity to the mean sea level by combining $\Psi$ with different $\Psi_i$ ($i = GGM, EGM$). It is expected that the definition of the mean sea level causes the dynamic system to respond differently to the forcing applied, that is, we expect to see different mean currents, and different (time scales of) variability.

2.1 Ocean model

The ocean model that we apply is the OPA model developed at LODYC [7]. We use the global configuration, ORCA2, with a 2-degree resolution in zonal direction, and whose meridional resolution is 2/3 degree at the equator. The model has 30 vertical levels, 20 of which lie in the upper 500 meters. Lateral mixing is performed along isopycnal surfaces, and an eddy parameterization scheme is used poleward of 20 degrees following [4]. Vertical mixing is modeled with the turbulent kinetic energy scheme of [2].

The model is forced with surface heat fluxes computed from bulk formulas, while the freshwater fluxes are represented by a relaxation of salinity to annual mean values from the [6] climatology. Climatological wind-stress fields over the period 1993-96 are obtained from a combination of ERS scatterometer winds [1] and wind-stress observations from the Tropical Atmosphere-Ocean (TAO) array [8].

2.2 Data assimilation scheme

Although successful analyses have been obtained with the four-dimensional variational scheme of OPA [14, 15], this method is relatively computer-time expensive. In addition, problems may occur as a result of the linearization in the tangent linear and adjoint models that are used in this scheme [13]. These problems are overcome when using ensemble methods, such as the Ensemble Kalman Smoother developed by [3]. This type of methods is especially suitable for nonlinear dynamics, and relatively easy to implement. In the current study, we apply the most straightforward implementation of ensemble methods, a smoother which we denote the direct ensemble method, following [11].

Because the exact derivation and statistical interpretation of the direct ensemble method can be found in [11], we limit the description here to the overall concept. A large number of model runs ($N$) is performed, each with slightly different forcing and/or initial condition, based on an a priori estimate of model error. At the end of a set of model simulations, the model output is compared to the observations. The analysis is then determined weighting each ensemble member $n = 1, ..., N$ with the probability density function of the observations, given the model realization. This probability density is assumed to be Gaussian and the resulting weighting function, $b_n$, is thus computed following Eq. 2:

$$b_n = \exp\left[-\sum_{i=0}^{M} \frac{\psi_{n,i} - d_i}{\sqrt{2}q} \right]^2,$$

(2)

where $\psi_{n,i}$ is the model realization of ensemble member $n$ at the location and time of the $i$th observation $d_i$, $q$ is the a priori observational error estimate, and $M$ the total number of observations. The weight factor for each ensemble member is normalized by the sum of all weight factors for all $N$ ensemble members.
2.3 Observations

As observational data set, we use T/P data processed at DEOS. Detailed information on the altimeter data processing can be found on the RADS web-site: http://www.deos.tudelft.nl/altim/rads/rads.shtml. Dynamic topographies have been created using different geoid models and an 8-year data set of T/P sea level observations. The data have been selected for the period May 1993- May 2001, computing a mean with respect to the EGM96 model [5] and corresponding anomalies. To obtain mean dynamic topographies with respect to other geoids, the differences of EGM96 with other models have been computed. As the other models are not as accurately known at high degrees/orders as EGM, they have been truncated at degree/order 120. The difference fields will thus still contain high-frequency signal from the EGM96 geoid. This has been removed by applying a Gaussian smoother on the geoid with a spatial scale (1-sigma) of 2.5 degree, which is similar to a cut-off at degree/order 70. The difference fields are then combined with the EGM96 mean dynamic topography to obtain the different dynamic topographies with respect to different geoid models. These dynamic topographies can be compared up to degree/order 70, which corresponds to a spatial resolution of 275 km. Considering the spatial resolution of the model (2 degrees), this is an acceptable scale. In the present paper, we only show the results obtained with the dynamic topography with respect to the GGM01C geoid model [10] (see also http://www.csr.utexas.edu/grace/gravity/).

2.4 Experiments

Each of the assimilation experiments consists of an ensemble of model simulations \( N = 128 \), which are compared to observations of dynamic height. To generate different ensemble members, we vary the initial conditions and the wind-stress forcing. The different initial conditions are obtained from a fifty-year model run by [9], each initial condition representing a different ocean state for January first of a particular year. The wind-stress variations for each ensemble are obtained by applying a random factor to the wind-stress fields (the climatological fields generated of ERS and TAO observations). Each ensemble member is forced with the same daily forcing, but the strength of the forcing is adapted in each model run by multiplying with a constant random factor that has mean 1 and standard deviation 1.

The formulation of the assimilation method in Section 2.2 allows assimilation of both sea level anomalies and mean sea level. In the present study however, we restrict ourselves to assimilation of the mean sea level. The experiment in which the dynamic topography with respect to EGM96 has been assimilated is called EnsEGM, the experiment with the dynamic topography with respect to GGM01C is denoted EnsGGM. The choice of \( q \) (in Eq. 2) in both experiments has been based on a comparison of different geoid models, and is set to 10 cm, constant throughout the basin. Although the global data sets are available, we concentrated on the equatorial oceans, and computed the ensemble weights based on a local analysis between 20\(^\circ\) N and 20\(^\circ\) S.

3 RESULTS

The ensemble model runs have root-mean-square (rms) differences in mean sea level around 30 cm, reaching highest amplitudes in regions of highest oceanic variability. The spread of the ensemble members is illustrated in Fig. 1. This rms difference of sea level in the ensemble should represent the model error in mean sea level, and is found to correspond reasonably well with the differences between different geoid models at spatial scales of 10 degrees and more.
By applying the weights of Eq. 2 to the different ensemble members, we obtain two different ocean analyses for EnsEGM and EnsGGM. The differences in sea level are depicted in Fig. 2. These differences are smaller than expected from the geoid difference as depicted in Fig. 3, but similar patterns are observed in both fields. Increasing the number of ensemble members will probably lead to a more pronounced difference between the ensemble-weights, and as a consequence larger differences between the analyses.

In the analyses, the most pronounced differences in sea level are observed in the equatorial region, where the meridional gradient in GGM01C is generally weaker than in EGM96, especially in the Northern Hemisphere. Even these small differences in sea level gradient (in the order of around 1 cm over 5 degrees), affect the strength of the equatorial undercurrent. Fig. 4 shows the difference in zonal current along the equator between the two weighted ensembles. The weakening of the meridional sea level gradient at the equator has resulted in a weakening of the Equatorial Undercurrent in both the Pacific and the Atlantic Oceans.

A striking difference in sea level occurs east of Madagascar. This difference is associated with a northward shift of the South Equatorial Current (result not shown).

4 DISCUSSION

The definition of the geoid affects the three-dimensional simulation of ocean circulation. The results above can only serve as a qualitative indication of this effect. To quantify this, the ensemble method presented above requires further elaboration. Although the variations in initial conditions and wind-stress forcing have shown to result in a large spread of the ensemble members, the solutions obtained for the two different dynamic topographies are so close that it is likely that the ensemble method has not yet converged. Therefore, the number of ensemble members should be increased, and the convergence of the method should be checked. The use of iterative importance resampling (a variation on sequential importance resampling) implies an effective selection of ensemble members, and is currently being investigated. Furthermore, using the perturbation to the wind-stress field, the spread of the ensemble is very much dominated by wind-driven dynamics and El Niño-like structures. This is by no means an optimal representation of the model error in mean sea level. The spread of the ensemble members can be further optimized by using a different perturbation method. We are currently investigating the use of stochastic optimals to disturb the most disruptive patterns of the atmospheric forcing.
Figure 4: Difference in zonal current along the equator between EnsEGM and EnsGGM. Contour interval is 0.5 cm/s.

(A.M. Moore, private communication). Once the method is effective, the length of the simulation should be increased to avoid interannual variations affecting the solution. As the dynamic topographies are constructed from 8 years of T/P observations, the corresponding model simulation should cover the same period.

Although immature, the ensemble method has the potential to identify the regions where the simulation of ocean circulation is sensitive to the definition of the geoid in a straightforward manner. The very preliminary study shown here illustrates this sensitivity, and makes us aware that accurate geoid information is essential for the simulation of ocean currents. Based on our findings, we expect GOCE to contribute significantly to our knowledge of the global ocean circulation.

References


