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Global eustatic sea-level variations for the approximation of geocenter motion from GRACE

Abstract: Global degree-1 coefficients are derived by means of the method by Swenson et al. (2008) from a model of ocean mass variability and RL05 GRACE monthly mean gravity fields. Since an ocean model consistent with the GRACE GSM fields is required to solely include eustatic sea-level variability which can be safely assumed to be globally homogeneous, it can be empirically derived from GRACE as well, thereby allowing to approximate geocenter motion entirely out of the GRACE monthly mean gravity fields. Numerical experiments with a decade-long model time-series reveal that the methodology is generally robust both with respect to potential errors in the atmospheric part of AOD1B and assumptions on global degree-1 coefficients for the eustatic sea-level model. Good correspondence of the GRACE RL05-based geocenter estimates with independent results let us conclude that this approximate method for the geocenter motion is well suited to be used for oceanographic and hydrological applications of regional mass variability from GRACE, where otherwise an important part of the signal would be omitted.

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1 Introduction

For about twelve years now, the Gravity Recovery and Climate Experiment (GRACE) satellite mission captures the time-variations in the Earth's gravity field at very large spatial scales. By assuming that low-frequency variations - i.e., at periods ranging from about 30 days to several years - take place entirely in a thin layer close to the surface of the solid Earth, those gravity field variations are unambiguously translated into mass anomalies (Wahr et al., 1998). These are primarily related to water re-distributions at and between the continents and the oceans.

Time-series of monthly mean gravity fields provided by the GRACE project are routinely calculated in a reference frame related to the center-of-mass of the Earth System (CM). The corresponding degree-1 terms of the spherical harmonic expansion are therefore zero by definition. Thus, mass anomalies for any regional average obtained from those series entirely neglect all contributions related to the offset between the center-of-mass and a center-of-figure (CF) frame, which is commonly denoted as geocenter motion (Petit and Luzum, 2010). In view of the high precision of GRACE estimates, however, geocenter motion contributions have been found non-negligible for both oceanographic (Chambers and Willis, 2009), and hydrologic applications (Chen et al., 2005).

Geocenter motion is typically derived from satellite laser ranging (SLR) observations to geodetic satellites as, e.g., LAGEOS and STELLA, that relate the kinematic orbits of those satellites to stations realizing the terrestrial reference frame (Chen et al., 1999; Eanes, 2000; Cretaux et al., 2002). Alternatively, observations of the global network of GPS permanent stations allow also for the derivation of geocenter variations (Blewitt et al., 2001; Lavallée et al., 2006; Fritsche et al., 2010). More recently, promising results of geocenter motion estimates have been reported for joint inversions of a wide number of different observations, i.e. satellite altimetry over the oceans, GPS permanent station observations, in-situ ocean bottom pressure observations, GRACE time-variable gravity fields (Davis et al., 2004; Jansen et al., 2009; Rietbroek et al., 2012). Partially augmented by information from geophysical models, those joint inversions provide geocenter motion estimates that are presumably more robust and less noisy than results from a single observing system alone.

Since geocenter motion estimates from all the methods mentioned above typically have latencies of several weeks to months, it would be highly beneficial to approximate the geocenter from data of the GRACE mission alone. A method allowing such an approximation has been proposed by Swenson et al. (2008). Their algorithm essentially assumes that from the a-priori knowledge of mass distribution at a sufficiently large fraction of the Earth's surface, the missing global degree-1 coefficients and thus the geocenter motion can be derived. The global oceans are the typical domain of choice, but other regions are in principle possible as well.

During the GRACE gravity field processing, high-frequency tidal and also non-tidal mass variability is de-aliased by means of time-variable background models (Flechtner and Dobslaw, 2013). The GRACE Level-2 monthly mean gravity fields - the so-called GSM fields - thereby do not include signals related to atmospheric mass variability. GSM fields also do not include ocean bottom pressure anomalies caused by regionally varying surface winds. They include however, signals related to the eustatic contribution to global sea-level, since this effect has been intentionally excluded from the de-aliasing model (Flechtner and Dobslaw, 2013). Note that the monthly-mean averages of both atmospheric and oceanic non-tidal variability removed during the de-aliasing process are provided as GAC products along with the GSM fields and therefore can be restored for further analysis if appropriate.

In order to approximate degree-1 coefficients consistent with the GRACE GSM fields with Swenson's method, a model of the eustatic contribution to global sea-level is required. Net-inflow of freshwater into the global ocean raises the sea-level homogeneously (Dobslaw and Thomas, 2007; Lorbacher et al., 2012) as long as effects of loading and self-attraction (Tamisiea, 2011; Kuhlmann et al., 2011) are not taken into account. Thus, such a model of the eustatic variation of the sea-level might be derived empirically from the GRACE gravity fields - by incorporating the assumption of a globally homogeneous response of sea-level to a net-flux of water.

The present paper is structured as follows: First, the essentials of the method of Swenson et al. (2008) are recalled in section 2. Next, we are going to test Swenson's method with the help of a numerical mass transport model developed for future satellite gravity mission simulation experiments. The model realistically includes mass redistributions among and within atmosphere, cryosphere, continental hydrosphere and the global oceans (section 3). Subsequently, we assess different processing strategies to empirically derive a model of the global eustatic sea-level variations from GRACE (section 4), that are finally applied iteratively to Swenson's method to arrive at a refined approximation of the geocenter motion that is finally discussed with respect to independent estimates from alternative geodetic techniques in section 5.

2 Methodology of Swenson et al. (2008)

The method to approximate geocenter motion introduced by Swenson et al. (2008) is essentially based on the fact that we can separate the surface mass signals into different components. The mass distribution at any position on a sphere is described from a (truncated) series of global spherical harmonic coefficients:

$$\Delta\sigma(\theta,\phi) = \sum_{l=0}^{\infty}\sum_{m=0}^{l} \tilde{P}_{lm}(\cos\theta)\{\Delta C_{lm}\cos m\phi + \Delta S_{lm}\sin m\phi\} \tag{1}$$

By setting a certain fraction of the sphere's surface - let's say all continental regions - to zero, a second series of spherical harmonic coefficients might be analyzed, that lead to identical mass distributions in the oceanic regions when re-synthesized:

$$\Delta C_{lm}^{ocean} = \frac{1}{4\pi}\int d\Omega\, \tilde{P}_{lm}(\cos\theta)\cos m\phi\; \vartheta(\theta,\phi)\;\sigma(\theta,\phi) \tag{2}$$

Here, ΔC_{lm}^{ocean} are called ocean coefficients in the remainder of this paper in order to distinguish them from the global coefficients ΔC_{lm}, and ϑ is a binary field separating land from ocean. By utilizing the orthogonality of the associated Legendre functions on the sphere, every single ocean coefficient might be analyzed individually:

$$\Delta C_{lm}^{ocean} = \frac{1}{4\pi}\int d\Omega\, \tilde{P}_{lm}(\cos\theta)\cos m\phi\; \vartheta(\theta,\phi) \cdot \sum_{l'=0}^{\infty}\sum_{m'=0}^{l} \tilde{P}_{l'm'}(\cos\theta)\{\Delta C_{l'm'}\cos m'\phi + \Delta S_{l'm'}\sin m'\phi\} \tag{3}$$

We now write the summation in Eq. (3) for the global coefficients of degree 1 explicitly:

$$\begin{aligned}\Delta C_{10}^{ocean} &= \Delta C_{10}\cdot\frac{1}{4\pi}\int d\Omega\,\tilde{P}_{10}(\cos\theta)\vartheta(\theta,\phi)\tilde{P}_{10}(\cos\theta)\\ &+\Delta C_{11}\cdot\frac{1}{4\pi}\int d\Omega\,\tilde{P}_{10}(\cos\theta)\vartheta(\theta,\phi)\tilde{P}_{11}(\cos\theta)\cos\phi\\ &+\Delta S_{11}\cdot\frac{1}{4\pi}\int d\Omega\,\tilde{P}_{10}(\cos\theta)\vartheta(\theta,\phi)\tilde{P}_{11}(\cos\theta)\sin\phi\\ &+\frac{1}{4\pi}\int d\Omega\,\tilde{P}_{10}(\cos\theta)\vartheta(\theta,\phi)\\ &\cdot\sum_{l=0}^{\infty}\sum_{m=0}^{l}\tilde{P}_{lm}(\cos\theta)\{\Delta C_{lm}\cos m\phi + \Delta S_{lm}\sin m\phi\}\end{aligned} \tag{4}$$

where the summation in Eq. (4) has to be read to exclude $\Delta C10$, $\Delta C11$, and $\Delta S11$. Equation (4) is re-expressed once more in a more compact form

$$\Delta C_{10}^{ocean} = \Delta C_{10}\cdot I_{10C}^{10C} + \Delta C_{11}\cdot I_{11C}^{10C} + \Delta S_{11}\cdot I_{11S}^{10C} + G_{10C} \tag{5}$$

where we use the notation

$$I_{10C}^{10C} = \frac{1}{4\pi} \int d\Omega \, \tilde{P}_{10}(\cos\theta)\vartheta(\theta,\phi)\tilde{P}_{10}(\cos\theta) \quad (6)$$

with the superscript indicating the spherical harmonic to the left of ϑ and the subscript indicating the one to the right, and

$$G_{10C} = \frac{1}{4\pi} \int d\Omega \, \tilde{P}_{10}(\cos\theta) \, \vartheta(\theta,\phi) \cdot \sum_{l=0}^{\infty} \sum_{m=0}^{l} \tilde{P}_{lm}(\cos\theta) \{\Delta C_{lm} \cos m\phi + \Delta S_{lm} \sin m\phi\} \quad (7)$$

Setting up corresponding equations for ΔC_{11}^{ocean} and ΔS_{11}^{ocean} leads to a linear equation system

$$\begin{bmatrix} \Delta C_{10}^{ocean} \\ \Delta C_{11}^{ocean} \\ \Delta S_{11}^{ocean} \end{bmatrix} = \begin{bmatrix} I_{10C}^{10C} & I_{11C}^{10C} & I_{11S}^{10C} \\ I_{10C}^{11C} & I_{11C}^{11C} & I_{11S}^{11C} \\ I_{10C}^{11S} & I_{11C}^{11S} & I_{11S}^{11S} \end{bmatrix} \begin{bmatrix} \Delta C_{10} \\ \Delta C_{11} \\ \Delta S_{11} \end{bmatrix} + \begin{bmatrix} G_{10C} \\ G_{11C} \\ G_{11S} \end{bmatrix} \quad (8)$$

that might be solved for ΔC_{10}, ΔC_{11}, and ΔS_{11} by matrix inversion in case that knowledge about the corresponding ocean coefficients at the left-hand side is available from auxiliary sources. For global GSM-like coefficients as explained above, this needs to be a semi-empirical model of the eustatic global sea-level variability (see section 4).

3 Numerical Experiments

Applying Swenson's methodology described above to obtain GSM-like global coefficients of degree-1 includes a number of potential sources of uncertainty: (i) The geometry of the fraction of the sphere where a priori information is introduced must be sufficiently large and sufficiently shaped to make the I matrix in Eq. (8) well conditioned and thereby numerically invertible. (ii) The GRACE GSM fields are obtained by reducing atmospheric and oceanic mass variability with the potentially erroneous time-variable background model AOD1B, thereby contributing to errors in the global coefficients contained in the G vector of Eq. (8). (iii) The empirical model of the eustatic sea-level rise based on limited observations might contain errors as well that are introduced into the algorithm via the left-hand side of Eq. (8).

Since the consequences of those effects are difficult to assess in general, we test the algorithm with four experiments with a decade-long model time series. For this, we use a numerical mass transport model developed for simulation studies of future satellite gravity missions. The numerical model is an update of the ESA Mass Transport Model (Gruber et al., 2011) and provides spherical harmonics of global mass anomalies complete to degree and order (d/o) 180 for the atmosphere based on ERA Interim (Dee et al., 2011), for the oceans based on OMCT (Dobslaw et al., 2013), for the continental hydrosphere based on LSDM (Dill and Dobslaw, 2013), and for the continental ice-sheets based on RACMO-2 (Ettema et al., 2009), all consistently forced with ERA-Interim atmospheric data. The time series cover twelve years from 1995 until 2006, and its temporal sampling is six hours. Since we are going to approximate the geocenter on time-scales corresponding to the nominal GRACE sampling, monthly mean averages calculated for each sub-system are used in the following experiments.

In TEST 1, we assume that the mass anomalies over the oceans are known exactly from the mass transport model. We take all coefficients of d/o = 0 and d/o ≥ 2 from the sum of ocean, continental hydrology and ice as the global coefficients, and derive degree-1 coefficients with Swenson's method that compare well to the true ones from the mass transport model (Fig. 1a). Relative explained variances of more than 99% for all components (Table 1) indicate that the geometry of the global ocean is well suited to derive global degree-1 terms with Swenson's algorithm.

For TEST 2, we take again the mass anomalies over the oceans as known, but additionally include the atmosphere into the global coeffcents (Fig. 1b). By this, we assume that none of the monthly-mean atmospheric contributions has been removed by the time-variable background model AOD1B, which is a quite conservative assumption, since current atmospheric models are generally found to remove a substantial amount of residual variability from the GRACE data (Zenner et al., 2012). Relative explained variances of more than 97% let us conclude that errors in the atmospheric component of the time-variable background model do not affect the methodology seriously, and therefore need not to be considered further in this context.

During TEST 3, we take the mass anomalies over the oceans from the mass transport model and calculate a time-series of the globally averaged eustatic sea-level changes. This eustatic sea-level contribution is subsequently spread out equally over the ocean domain of the mass transport model and used together with the global coefficients of TEST 1 to calculate once more the global degree-1 coefficients (Fig. 1c). With this experiment, we find substantial deviations between true and recovered global degree-1 coefficients, notably in ΔC_{11}, where only 42% of the variance in the true coefficient is described by the one obtained with Swenson's algorithm and a globally homogeneous model of eustatic sea-level variability. In

Table 1. Relative explained variance in [%] for the recovery of global degree-1 coefficients in four different test calculations with the updated ESA mass transport model.

	ΔC_{10}	ΔC_{11}	ΔS_{11}
TEST 1	99.48	99.90	99.62
TEST 2	97.21	98.32	99.17
TEST 3	81.79	42.29	76.74
TEST 4	82.28	42.37	76.42

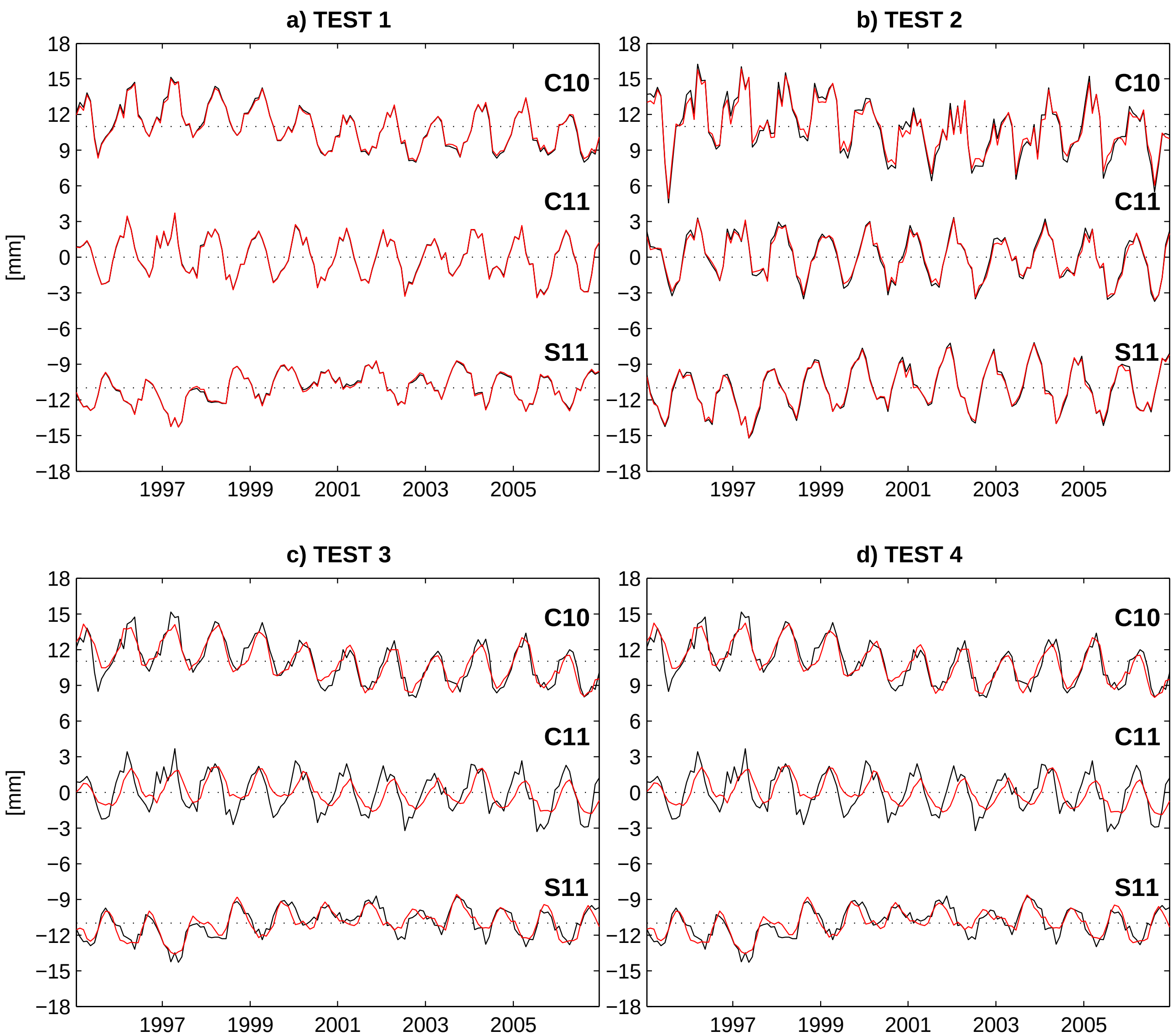

Fig. 1. Recovery of global geocenter variations in four different test calculations with the updated ESA mass transport model: true model series [black] and estimated with the method of Swenson et al. (2008) [red].

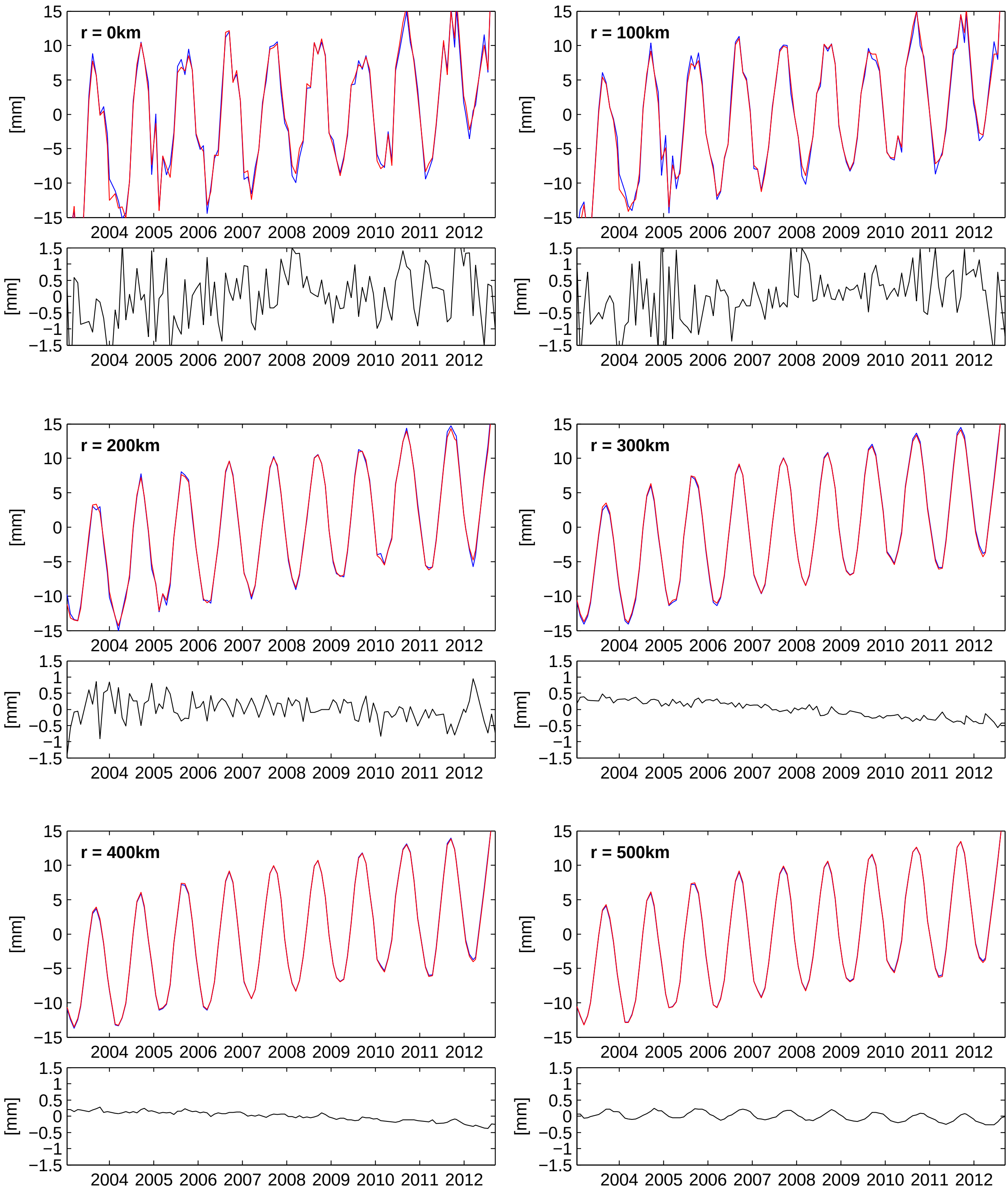

Fig. 2. Global ocean mass variations estimated with Gauss Filter [red] and Langrange-Multiplier-Method [blue] for different filter radii (r = 0, 100, ... , 500 km). The differences between both methods (red - blue) are given in the subplot below each plot.

particular the seasonal variability is affected, with an apparent phase-shift of several weeks. Secular trends, however, are captured well even in this experiment.

Finally, for TEST 4, we repeat the previous experiment but scale the homogeneous model of eustatic sea-level variability with a factor of 1.1 in order to assess the impact of potential errors in the global sea-level model (Fig. 1d). By comparing derived global degree 1 coefficients from experiments 3 and 4, we find scaling coefficients of 1.031 for ΔC_{10}, 1.036 for ΔC_{11}, and 0.983 for ΔS_{11}, suggesting that in particular ΔC_{11} and ΔC_{10} are susceptible to errors in the eustatic sea-level model. We will return to those results in the remainder of this paper.

4 Global eustatic mass variations from GRACE

As outlined above, a model of eustatic sea-level variability in the global oceans is required to derive global GSM-like coefficients with Swenson's method. To obtain one, we follow the approach of Chambers (2004) that is entirely based on GRACE monthly mean gravity fields.

We take the release 05 (RL05) GRACE gravity field solutions of the Deutsches GeoForschungsZentrum (GFZ) (Dahle et al., 2012) provided as Stokes coefficients up to degree/order 90 for the period February 2003 to September 2012. We introduce an annual sinusoid for global degree-1 coefficients taken from Eanes (2000). In order to minimize continental leakage, we use an averaging domain that excludes all ocean areas closer than 300 km to the coasts. Averaging mass variability over the chosen domain has been performed in two different ways as proposed by Swenson and Wahr (2002): (i) an isotropic Gauss filter, and (ii) the Lagrange Multiplier Method. Where the former method attempts a compromise between satellite and leakage error, the latter will fit the leakage error to satellite errors obtained from the GRACE formal error. Note that due to the large averaging area applied here, the consideration of anisotropic filter algorithms as suggested by Kusche (2007) is not necessary, since it does not notably affect the results. Signal loss, due to the filtering, however, is taken into account by applying re-scaling factors following Klees et al. (2007).

For both the Gauss filtering and the Lagrange Multiplier Method, we apply smoothing radii between 0 and 500 km, the corresponding re-scaling factors vary between 1.002 to 1.014, and 1.027 to 1.046, respectively. Amplitudes (Table 2) compare well between the two methods, and generally decrease with stronger smoothing, as expected. Formal uncertainties of the amplitudes do not decrease substantially for smoothing radii above 300 km. Differences between the time-series of both methods (Fig. 2) are small for 300 km, and show - apart from a slightly differing secular trend - no systematics. Since over-smoothing tends to diminish the signal and increases the chance of continental leakage, and since 300 km is also close to the theoretical spatial resolution limit of GRACE that is determined by the inter-satellite distance of 250 km, and since moreover both methodologies lead to identical amplitudes within their formal uncertainties, we decide to use the 300 km Gauss filtered results for subsequent analysis.

5 GSM-like global degree-1 coefficients from GRACE

We now introduce the empirical model of eustatic sea-level variability as a apriori information into Swenson's method for the derivation of GSM-like global degree-1 coefficients. For the G vector of Eq. (8) we use again GRACE RL05 gravity field solutions from GFZ (Dahle et al., 2012) up to d/o= 90. No additional filtering is applied to the GRACE coefficients at this stage, but to remove the long-term signal of the Glacial Isostatic Adjustment (GIA), we subtract the model of Paulson et al. (2007).

We estimate global degree-1 coefficients for the empirical eustatic sea-level model with the methodology of Swenson (Fig. 3) and compare them to coefficients based on CSR RL05 solutions from GRACE that are readily available from the GRACE TELLUS website (http://grace.jpl.nasa.gov/data/degree1) maintained by Sean Swenson. In general, we note good agreement between both solutions, in particular with respect to secular changes and annual amplitudes. Small month-to-month differences exist, however, which are related to the different GRACE series applied, and to the specifics of the processing of the eustatic sea-level variability model. In addition, we note stronger deviations between both solutions after 2011, when GFZ RL05 solutions were found to be overly constrained to secular trends in a time-variable low-degree background model, which eventually led to the replacement of RL05 with RL05a in autumn 2013 (see Release Note for GFZ RL05 GRACE L2 Products, C. Dahle, 2014, isdc.gfz-potsdam.de).

We now return to the fact that we introduced a priori information on the global degree-1 coefficients from Eanes (2000) into the eustatic sea-level variability model, which in principle can be replaced by the newly derived global degree-1 coefficients as given in Fig. 3. We there-

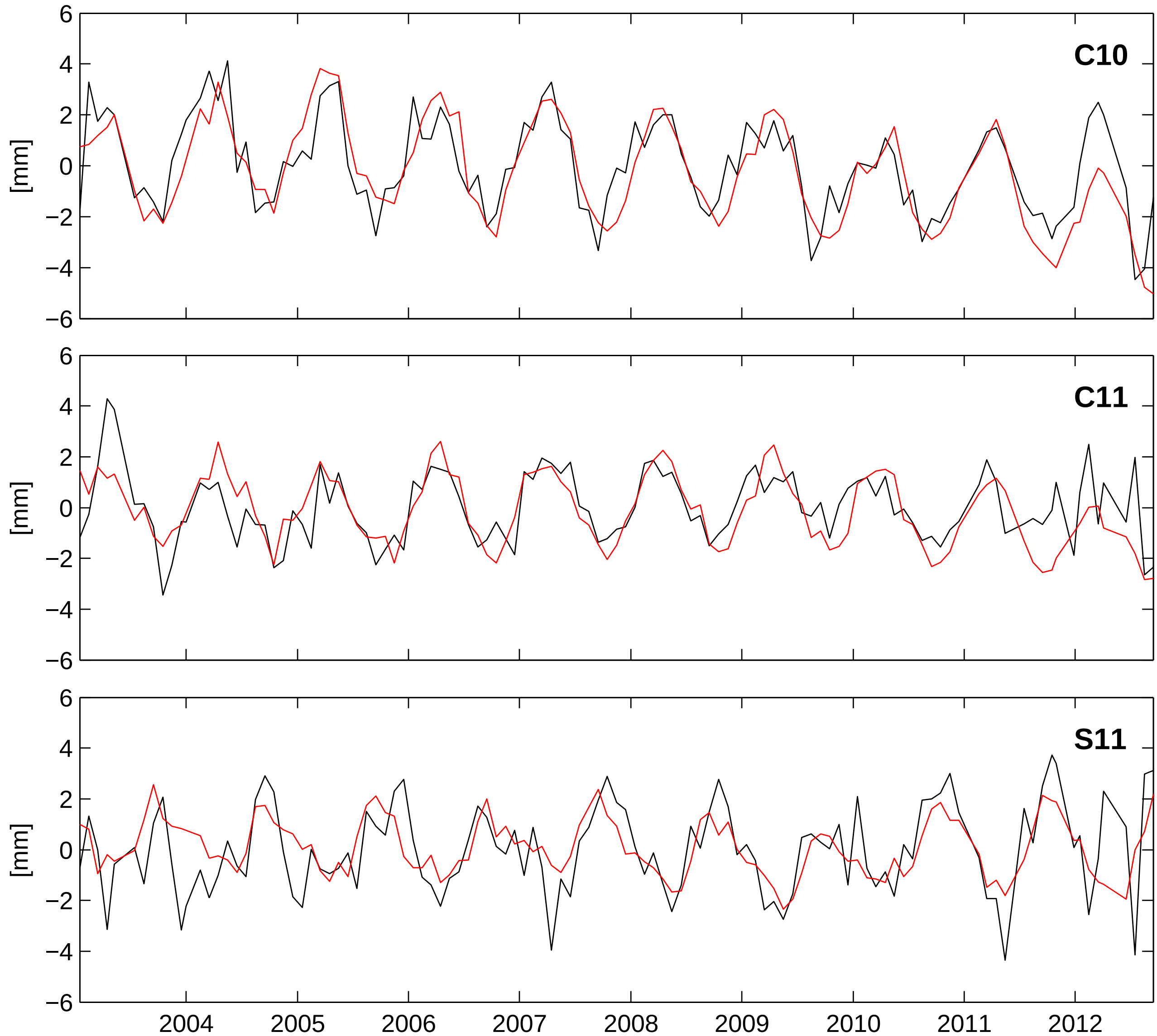

Fig. 3. Estimated global geocenter variation time series from GFZ RL05 [black] with global ocean mass variations estimated with Gauss Filter for a filter radius of 300 km, and provided by the GRACE TELLUS webpage (http://grace.jpl.nasa.gov/data/degree1/) for the CSR RL05 solutions [red].

fore repeat the calculations for the global model of eustatic sea-level variability, re-do the degree-1 estimation with Swenson's method, and iterate five times. Decreasing increments during those iterations indicate good convergence of the method (Fig. 4). In addition to this reference calculation, we perform an experiment 1 by setting the global degree-1 coefficients to zero in the initial calculation of the eustatic sea-level variability model, perform the degree-1 calculation based on this model, and perform again five iterations. For another experiment 2, we start with ocean degree-1 coefficients set to zero in Swenson's method, use the derived first estimate of the global coefficients as input for the eustatic sea-level model estimation, and iterate again for 5 times. For all three cases (reference as well as both experiments), increments decrease by approximately one order of magnitude per iteration, indicating that the initial assumption on the degree-1 terms introduced does not have a notable effect.

The final estimate of eustatic sea-level variability compares favorably with previously published results (Table 3). Amplitudes are slightly larger than early GRACE results from Chambers (2004), which might be explained by the fact that those early results have been obtained from the very first GRACE release with substantially higher error

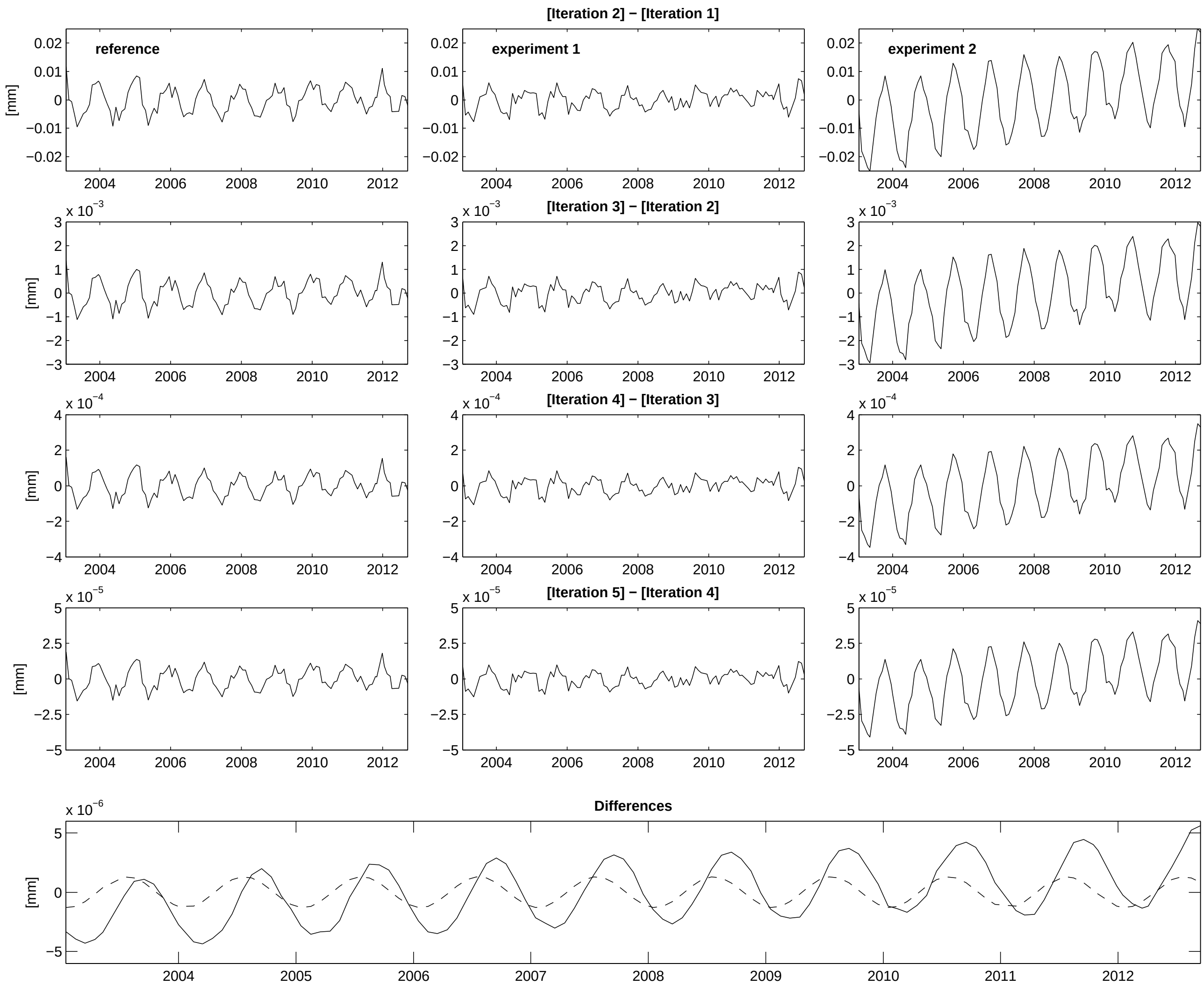

Fig. 4. Increments in eustatic sea-level from the iterative determination of global degree-1 coefficients and a globally homogeneous eustatic sea-level variability model: starting from global degree-1 coefficients given by Eanes (2000) (reference, left column), starting from zero global degree-1 coefficients (experiment 1, middle), starting from zero eustatic sea-level anomaly (experiment 2, right). The bottom figure shows the difference of the time series after 5 iterations for reference minus experiment 1 [solid line] and reference minus experiment 2 [dashed line].

levels and therefore smoothing requirements. More recent calculations as published by Wouters et al. (2011) correspond quite closely to our results. Removing the seasonal cycle from the eustatic sea-level curve discloses for example the drop in sea-level in 2011 (Fig. 5), that has been related to record-high precipitation rates in Australia caused by La Nina teleconnections in the Central Pacific region (Fasullo et al., 2013).

The obtained GSM-like degree-1 coefficients are finally compared to independent results from a joint inversion (Rietbroek et al., 2012) (Fig. 6a). In general, we note higher short-term variability in the joint inversion, which is certainly due to the shorter time-sampling of only seven days compared to the 30 day averages considered in this paper. It is however, interesting to note that the joint inversion apparently follows the AOD1B RL05 degree-1 coefficients quite closely (gray line in Fig. 6a), even though the product is stated to be GSM-like. For the full geocenter solutions, that are obtained in our case by adding the GSM-like solutions to AOD1B RL05 averages as provided with the GRACE gravity field by means of GAC products, this phase-shift between the two solutions is no longer visible. Here, geocenter motion as approximated by the method of Swenson corresponds well to the joint inversion, whereas a solution solely based on SLR observations deviates much stronger in particular in the ΔC_{10} and ΔC_{11} components.

Table 2. Annual amplitude and phase of global ocean mass variations determined from GRACE GFZ RL05 with a) Gauss Filter and b) Lagrange-Multiplier-Method for different filter radii.

	Amplitude [mm]	Phase [days]
a) Gauss Filter		
r = 0km	10.33±2.28	257
r = 100km	10.03±0.80	257
r = 200km	9.69±0.18	255
r = 300km	9.53±0.12	255
r = 400km	9.45±0.11	254
r = 500km	9.37±0.10	254
b) Lagrange-Multiplier-Method		
r = 0km	10.37±2.30	257
r = 100km	10.25±1.11	257
r = 200km	9.77±0.24	255
r = 300km	9.57±0.13	255
r = 400km	9.41±0.11	254
r = 500km	9.26±0.10	254

Table 3. Annual amplitude and phase of global ocean mass variations.

	Amplitude [mm]	Phase [days]
a) final results after 5 iterations		
this study	9.78±0.54	278
b) other studies		
Chambers (2004)	8.40±1.1	266
Rietbroek et al. (2009)	8.70	247
Siegismund et al. (2011)	8.10	252
Wouters et al. (2011)	9.40	280
Hughes et al. (2012)	8.12	266

Table 4. Annual amplitude and phase of finally estimated GSM-like and full (GSM+GAC) geocenter variations.

	$\Delta C10$		$\Delta C11$		$\Delta S11$	
	[mm]	[days]	[mm]	[days]	[mm]	[days]
a) GSM-like coefficients						
this study	2.04±0.18	88	1.39±0.19	105	1.54±0.34	−84
Rietbroek et al. (2012)*	3.01±0.37	20	1.94±0.14	63	3.47±0.11	−39
b) full geocenter coefficients						
this study	2.15±0.42	63	1.93±0.31	59	2.52±0.34	−45
Cheng et al. (2010)*	4.37±1.78	39	4.17±0.90	30	2.62±1.29	−39
Rietbroek et al. (2012)*	3.09±0.39	20	1.99±0.16	63	3.56±0.11	−38

* The mean, trend and annual signal component has been fitted to the time series of Cheng et al. (2010) and Rietbroek et al. (2012) and the formal errors of the fits have been estimated.

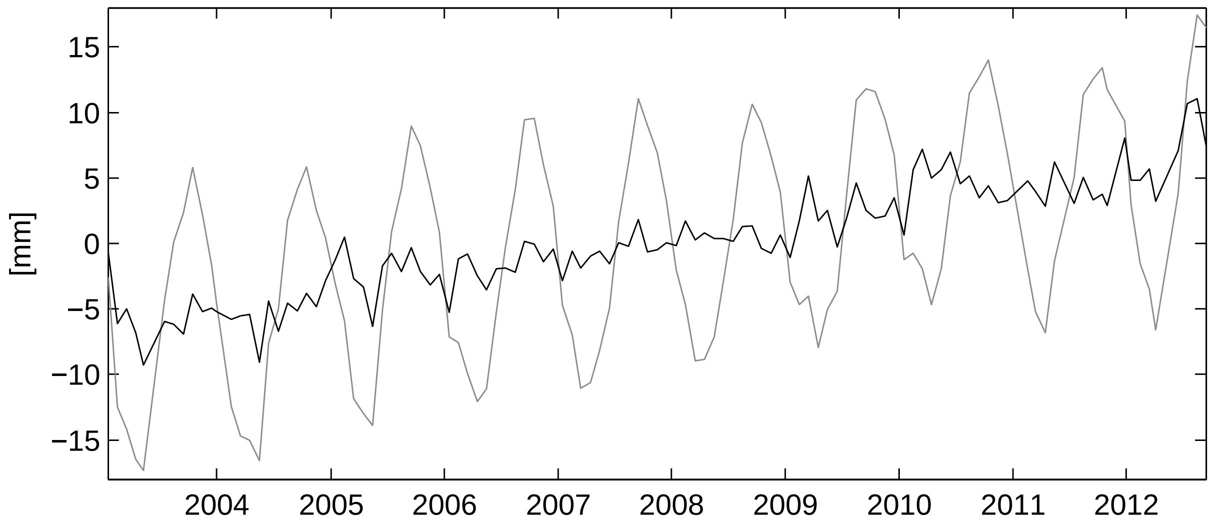

Fig. 5. Final estimated global ocean mass variations after 5 iterations [gray], and global ocean mass variations after 5 iterations with the annual cycle removed [black].

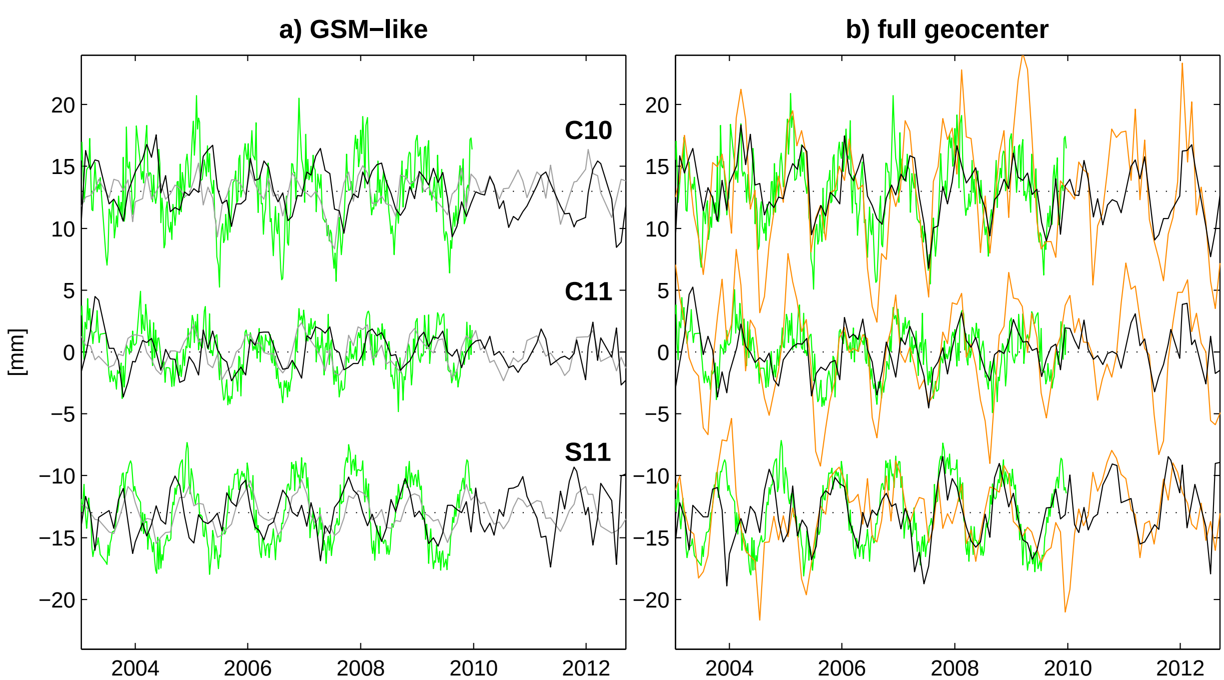

Fig. 6. Left: GSM-like geocenter variations from GRACE with global ocean mass variations with Gauss Filter (300km, 5 iterations) [black], results from joint-inversion by Rietbroek et al. (2012) [green] and degree-1 terms of the RL05-GAC product [gray]. Right: full geocenter time series (RL05 GSM + GAC) from GRACE with global ocean mass variations estimated with Gauss Filter (300 km, 5 iterations) [black], geocenter variations from Satellite Laser Ranging measurements from Cheng et al. (2010) [orange] and results from joint-inversion by Rietbroek et al. (2012) [green].

6 Summary and Conclusions

A methodology to approximate geocenter motion that was originally proposed by Swenson et al. (2008) essentially requires (i) a truncated series of global Stokes coefficients of degree two and higher, and (ii) degree-1 terms of the ocean bottom pressure variability. Since atmosphere and ocean dynamics are routinely removed from the GRACE gravity fields, global degree-1 terms consistent with the GRACE GSM fields should contain over the oceans only contributions from eustatic sea-level variability. By assuming that the eustatic sea-level change is globally homogeneous, such a model might be obtained from a truncated series of global Stokes coefficients of degree two and higher. Under those assumptions it is therefore possible to approximate GSM-like geocenter motion solely from GRACE monthly mean gravity fields.

Our implementation of the methodology of Swenson et al. (2008) has been thoroughly tested: experiments with a 12 year-long model series indicate that errors in the atmospheric part of the AOD1B that might potentially leak into the GSM fields can be safely neglected. Errors in the ocean component are potentially more severe, and affect in particular the ΔC_{11} component. Moreover, we note a sensitivity of the method to uncertainties in the eustatic sea-level model in particular for the ΔC_{10} and ΔC_{11} components. Initial assumptions on global degree-1 terms, however, are not important, since convergence is always reached after less than five iterations in every single test case considered.

The geocenter motion estimates from GRACE compare well to both GSM-like coefficients from a joint inversion and - after adding back the corresponding GAC field of monthly mean atmosphere and ocean mass variability removed during the gravity field processing by means of the AOD1B background model - to full geocenter estimates based on SLR. It is interesting to note that correspondence with SLR is generally best for ΔS_{11}, the component that is less susceptible to errors in the eustatic sea-level model as shown in section 3. Also, the close correspondence of the GSM-like results of the joint inversion with the AOD1B-only geocenter estimates, especially in seasonal frequencies, might be worth to be analyzed in more detail in the future.

For regional oceanographic and hydrologic applications of GRACE satellite gravimetry fields, however, the generally good correspondence of the approximated geocenter motion estimates from GRACE with independent solutions let us conclude that this data-set is well suited to be used where otherwise an important part of the signals would be missed out. Since all information required for the algorithm is available from the GRACE archives, this would allow for another step towards a near real-time monitoring of mass distribution and transport in the Earth system by means of satellite gravimetry.

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