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Evaluation of Release-05 GRACE time-variable gravity coefficients over the ocean

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Abstract. The latest release of GRACE (Gravity Recovery and Climate Experiment) gravity field coefficients (Release-05, or RL05) are evaluated for ocean applications. Data have been processed using the current methodology for Release-04 (RL04) coefficients, and have been compared to output from two different ocean models. Results indicate that RL05 data from the three Science Data Centers – the Center for Space Research (CSR), GeoForschungsZentrum (GFZ), and Jet Propulsion Laboratory (JPL) – are more consistent among themselves than the previous RL04 data. Moreover, the variance of residuals with the output of an ocean model is 50–60 % lower for RL05 data than for RL04 data. A more optimized destriping algorithm is also tested, which improves the results slightly. By comparing the GRACE maps with two different ocean models, we can better estimate the uncertainty in the RL05 maps. We find the standard error to be about 1 cm (equivalent water thickness) in the low- and mid-latitudes, and between 1.5 and 2 cm in the polar and subpolar oceans, which is comparable to estimated uncertainty for the output from the ocean models.

1 Introduction

Several versions of mapped ocean bottom pressure (OBP) anomalies determined from GRACE (Gravity Recovery and Climate Experiment) time-variable gravity coefficients have been provided to the scientific community via the GRACE *Tellus* website at Jet Propulsion Laboratory (JPL) (http://grace.jpl.nasa.gov/) from the two central GRACE Science Data System (SDS) centers (the Center for Space Research (CSR) and Helmholtz Centre Potsdam GFZ German Research Centre for Geosciences (GFZ)). The grids are based on Release-04 (RL04) coefficients (Bettadpur, 2007) and are post-processed to reduce correlated errors which appear as north–south stripes in the data (Swenson and Wahr, 2006), using either an ad hoc destriping algorithm and additional Gaussian smoothing that was optimized for the ocean (Chambers, 2006), or by projecting GRACE data onto empirical orthogonal functions (EOFs) from an ocean model (Chambers and Willis, 2010). Uncertainty of the mapped data has been estimated to be between 2 and 3 cm root-mean-square (RMS) depending on the type of processing, based on comparison to steric-corrected altimetry (Chambers, 2006; Chambers and Willis, 2010), output from an ocean model (Ponte et al., 2007; Quinn and Ponte, 2010), or bottom pressure recorders (Morison et al., 2007; Park et al., 2008). Even with this level of uncertainty, however, the mapped OBP data from GRACE have proven useful in a number of studies as disparate as, for example, studying low-frequency changes in the Arctic (Morison et al., 2007) or the North Pacific (Chambers and Willis, 2008; Song and Zlotnicki, 2008; Chambers, 2011), variability of the Antarctic Circumpolar Current transport (Zlotnicki et al., 2007; Boening et al., 2010; Bergmann and Dobslaw, 2012), exchange of mass between basins (Ponte and Quinn, 2009; Chambers and Willis, 2009), or record anomalous pressure changes in the Southern Ocean (Boening et al., 2011).

Recently, CSR, GFZ, and JPL have all produced a new release of GRACE gravity field coefficients – designated Release-05 (RL05) – spanning six years, from January 2005 to December 2010 (Bettadpur, 2012; Dahle et al., 2012). Major changes between RL04 and RL05 include improved knowledge of alignments between the star camera, accelerometer, and K-band ranging system for Level-1B data, and updated mean gravity field, ocean tide, pole tide, and

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de-aliasing models for Level-2 processing. The goal of this article is to analyze these new GRACE data over the ocean in a similar fashion as the older data and quantify the improvement of RL05 data over RL04. This will be done by comparing GRACE OBP with that output from two different ocean models. This should give a more accurate uncertainty estimate than comparing to OBP derived from altimetry corrected for either seasonal steric variations (Chambers, 2006) or monthly steric fluctuations (Chambers and Willis, 2010), as these data tend to have unresolved interannual steric fluctuations and/or sample mesoscale eddies that are far larger than OBP variations. Moreover, recent studies comparing RL04 GRACE data to ocean models have shown better agreement than earlier studies (e.g., Quinn and Ponte, 2010), suggesting models are now producing more reasonable low-frequency OBP variations. By comparing GRACE to two different models and by comparing the models to one another, we will show a method to estimate uncertainty in the GRACE maps as well as those from the models.

After demonstrating that the RL05 data are significantly improved over the RL04 data using the same post-processing methods, we will further investigate whether the parameters of the destriping algorithm can be relaxed for the RL05 data and still allow for similar or lower uncertainty. Section 2 will give an overview of the post-processing steps and ocean models used for the analysis, Sect. 3 will present the comparison between RL04 and RL05 maps and the error analysis, Sect. 4 will examine whether the destriping algorithm can be relaxed without increasing uncertainty, and Sect. 5 will summarize the optimal processing of GRACE RL05 gravity data when examining ocean bottom pressure.

2 Review of Release-04 data processing and ocean models

Details of the post-processing of released GRACE gravity coefficients to produce de-striped coefficients for ocean applications, and converting from gravity coefficients to mapped OBP in terms of equivalent sea level can be found in Chambers (2006) and Chambers and Schröter (2011). Here, we will review steps and point out improvements that are used in the current RL04 post-processing, implemented after Chambers (2006).

First, a long-term mean gravity field is removed from the coefficients to remove the time-invariant signal related to the solid earth gravity. This is done by averaging the monthly gravity coefficients reported by CSR, GFZ, and JPL between January 2005 and December 2010, and removing the mean coefficient from each month to compute anomalies. The monthly degree 2, order 0 coefficients estimated with GRACE are replaced with those from a satellite laser ranging analysis (Cheng and Tapley, 2004), due to significant errors in observing that coefficient with GRACE. Monthly geocenter estimates based on the method of Swenson et al. (2008) have been applied, as GRACE does not detect these. The method is based on a combination of GRACE gravity coefficients over the land and ice sheets and OBP from a model, including mean ocean mass variability. A correction for glacial isostatic adjustment (GIA) has been applied based on the model by Paulson et al. (2007), in order to remove the secular trend in the gravity field that is not due to the recent redistribution of water over the Earth's surface (Chambers et al., 2010).

The GRACE coefficients have correlated errors that map into vertical stripes as first described by Swenson and Wahr (2006). In Chambers (2006), we modified the algorithm proposed by Swenson and Wahr (2006) to make it more applicable to the longer-wavelength, small amplitude OBP variations and tested it on RL02 data. In processing the RL04 data, we found that several parameters in the filter could be relaxed and still provide maps without significant stripes. The filter that has been implemented for RL04 coefficients is generally the same as the one described in Chambers (2006) except that it keeps the lower 11×11 portion of the coefficients unchanged (increased from the lower 7×7) as well as all order 0 and order 1 coefficients. A 5th order polynomial is fit as a function of even or odd degree (n) to the remaining coefficients (reduced from a 7th order polynomial) for each order (m) greater than 2 from $n = 12$ (or $n = m$ if $m > 11$) up to $n = 60$. Only one polynomial is computed for each odd or even set for a given order unlike the method of Swenson and Wahr (2006), which calculates multiple polynomials for each series as a running computation. Only coefficients up to $m = 40$ are de-striped. Every coefficient above $n = 40$, $m = 40$ is set to zero (reduced from n=80). The maximum order has been reduced from the filter in Chambers (2006) because CSR_RL04 coefficients were only solved to $n = 60$, $m = 60$, and there is little difference in OBP over the open ocean from a model truncated at degree/order 40 compared to full resolution, provided the data are not further smoothed with a Gaussian with a radius longer than approximately 500 km (Fig. 1). Although differences in shallow water are larger due to shorter-wavelength barotropic fluctuations, GRACE will have problems observing these due to leakage of much larger land hydrology fluctuations, so the algorithm is optimized for finding long-wavelength open ocean OBP variations. The filter is applied to the coefficients of each month and each center separately, as the correlated errors differ from month-to-month and center-to-center.

Monthly averages of the modeled non-tidal ocean bottom pressure variations (available as GAD files from each processing center) are added back to the GRACE coefficients in order to return the full, monthly OBP variation, since GRACE gravity coefficients are estimated relative to this model. Note that this model is primarily designed to model high-frequency OBP variations that will alias into the short-wavelength gravity field. Low-frequency, long-wavelength errors in the model will be corrected by the GRACE estimation, so that when the monthly background model is restored,

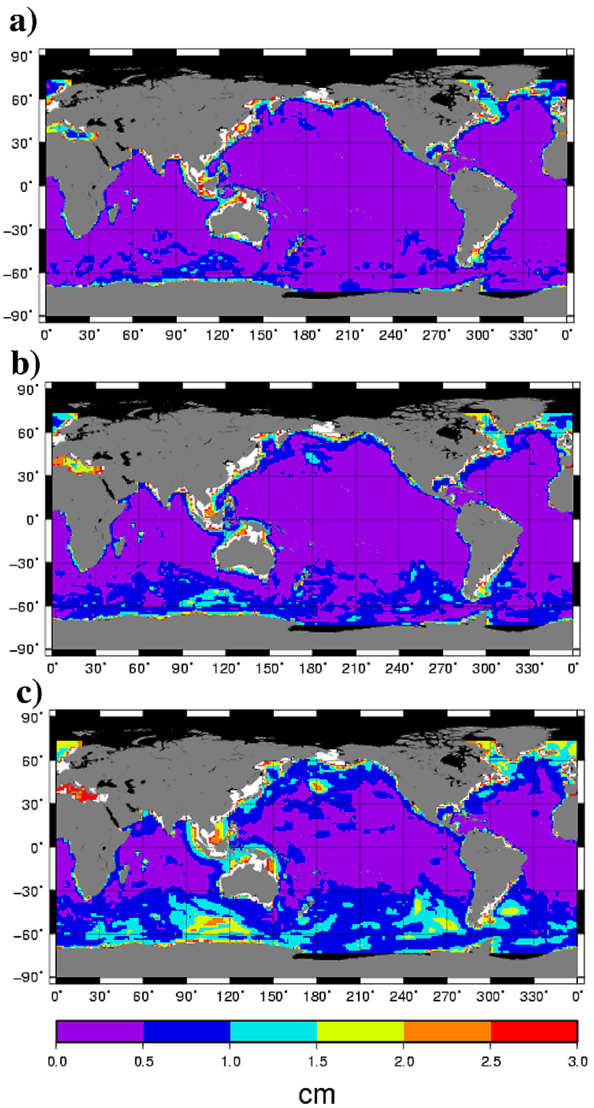

Fig. 1. Standard deviation of differences between unsmoothed OBP from JPL_ECCO and **(a)** JPL_ECCO OBP truncated to spherical harmonic degree/order 40 and smoothed with a 300 km Gaussian, **(b)** JPL_ECCO OBP truncated to degree/order 40 and smoothed with a 500 km Gaussian, and **(c)** JPL_ECCO OBP truncated to degree/order 40 and smoothed with a 750 km Gaussian.

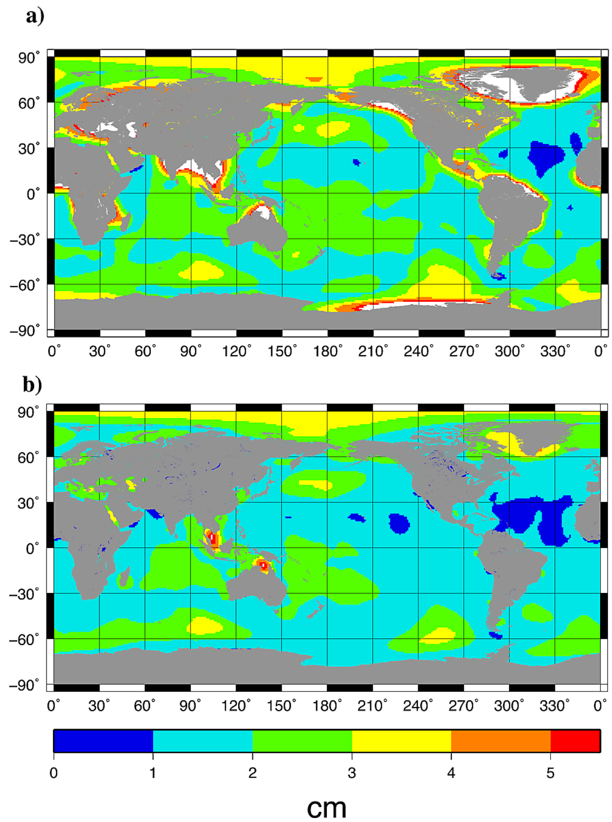

Fig. 2. Standard deviation of OBP from GRACE (CSR_RL05) **(a)** without leakage correction and **(b)** with leakage correction.

the combination reflects the unaliased monthly OBP that would have been sensed by GRACE if no model was used (Chambers and Willis, 2009).

Due to the large-scale smoothing used to extract the small amplitude OBP variations, larger variations from land hydrology and ice mass loss will leak into the ocean near the land-ocean boundary, extending out by about 500 km (Fig. 2). In Chambers (2006), we simply masked these data out. For the Release-04 processing, however, we used a method first proposed by Wahr et al. (1998) to use GRACE observations over land as a model of the land mass variability to compute and remove the leaked signal. We could have used output from a land hydrology model, but this has several deficiencies. First, no global hydrology model contains the mass losses from the ice sheets or glaciers, which are now the largest mass fluctuations in the GRACE observations. Second, hydrology models tend to model soil moisture and snow fairly well, but not surface water or groundwater. Since GRACE will observe both the ice losses and combined hydrology variations, it provides a better estimate of the signals than just a hydrology model. To compute the leaked signals, we first compute the gridded mass densities from GRACE data with no filtering. Then we mask out ocean areas and convert the land-only mass variations back to gravity coefficients. These are then destriped and smoothed identically to the processing used to compute the OBP maps, and the values over the ocean are subtracted to remove the leaked signal. The method removes the majority of the leakage around continents (Fig. 2), although there is some residual leakage left around Greenland and the Alaskan glaciers that even this method cannot fully reduce.

Two general ocean circulation models are also used in the analysis. The first is a version of the MIT general circulation model (Marshall et al., 1997) that is run at JPL as part of the Estimating the Circulation and Climate of the Ocean

(ECCO) consortium. We use monthly values of ocean bottom pressure derived from version kf080 that is available at http://grace.jpl.nasa.gov. This version of JPL_ECCO is a baroclinic model forced by winds, pressure, and heat and freshwater fluxes from the National Center for Environmental Prediction (NCEP) operational analyzes products and also assimilates satellite altimetry (Fukumori, 2002; Kim et al., 2007). The JPL_ECCO model extends only between $\pm 78°$ latitude; therefore, does not model OBP fluctuations in the Arctic Ocean or near Antarctica. Large differences between GRACE and JPL_ECCO at these boundaries should probably be considered errors in the model because of this limitation.

The GRACE project uses output from the Ocean Model for Circulation and Tides (OMCT) to compute high-frequency OBP in order to de-alias GRACE data during processing (Thomas, 2002; Flechtner, 2007). Similar to ECCO, this is a baroclinic model forced by winds, pressure, and heat and freshwater fluxes from the ECMWF operational analyzes. Unlike JPL_ECCO, OMCT models the entire ocean, including the Arctic and Antarctic. The monthly average OBP from this model (combined with overlying atmospheric pressure and estimated only over days when GRACE data was available) is distributed as GAD files along with the GRACE gravity coefficients. There are significant differences between the RL04 version of the OMCT model and the RL05 version, mainly to improve resolution and incorporate changes in parameterization that allowed better matches with in situ observations not available when the original version was held fixed for GRACE processing. One aspect that has improved is the high-frequency variability, which is important for dealiasing. The RL04 version has been shown to have significant deficiencies at periods less than a month in two recent studies (Bonin and Chambers, 2011; Quinn and Ponte, 2011, 2012). For this analysis, we will use the RL05 version, based on the distributed GAD files.

Neither ECCO nor OMCT model the time-variable global mean fluctuation in OBP caused by the exchange of water mass among the land, ocean, and atmosphere which GRACE does measure (e.g., Chambers et al., 2004, 2010). If this difference was not accounted for, the RMS of differences would be biased high due to missing this nearly 1 cm annual signal. In order to make the models consistent with the GRACE observed OBP, we add the GRACE observation of monthly mean ocean mass to the JPL_ECCO and OMCT grids before computing statistics. Moreover, JPL_ECCO does not contain the time-variable global mean fluctuation in OBP due to changes in the mean atmospheric pressure over the ocean (e.g., Ponte et al., 2007), which is a 0.6 cm seasonal variation. This is included in the GAD coefficients (and so is also included in the GRACE observations). To make data consistent, we add the monthly mean pressure from the GAD data to JPL_ECCO.

3 Analysis of Release-05 data

The new RL05 coefficients were initially processed exactly as the RL04 coefficients described in Sect. 2, with the exception that the geocenter estimates are based on RL05 GRACE gravity data combined with RL05 Atmosphere-Ocean Dealiasing (AOD) OBP from the GAD files using the method described in Swenson et al. (2008). The $C_{2,0}$ coefficients in the GFZ_RL05 solutions are considerably closer to the SLR estimates than either the CSR_RL05 or JPL_RL05 solutions, likely because GFZ uses a background time-variable gravity model based on RL04 coefficients where the $C_{2,0}$ value had been replaced with that from SLR. We tested statistics with and without replacing the $C_{2,0}$ coefficient in the GFZ_RL05 data, and found they were not significantly better. Since replacing the $C_{2,0}$ coefficient is still required for CSR_RL05 and JPL_RL05, we chose to replace the coefficient for consistency.

To demonstrate the reduced uncertainty in the RL05 data, the JPL_ECCO OBP maps (unsmoothed) are subtracted from the destriped and 300 km smoothed RL04 and RL05 OBP maps and the standard deviation of the residuals are computed (Fig. 3). Note that for the rest of the analysis, we mask out areas within 500 km of coastlines. This is to focus attention on the deep ocean where OBP variations are longer-wavelength and more resolvable by GRACE, and to quantify accurate statistics for the deeper ocean areas that are unbiased by higher errors near the coast. Coastal regions have very large, short-wavelength signals related to baroclinic interactions with the shelves, which are present to some extent in the JPL_ECCO data (Fig. 1), but will never be resolvable in GRACE. Additionally, even with the leakage correction described in Sect. 2, GRACE data still have higher uncertainty in near-coastal waters.

The improvement in the RL05 maps is obvious. The standard deviation of RL05 residuals is generally less than 2 cm throughout the ocean, and often less than 1.5 cm. Compare that with RL04 residuals, where the standard deviation is generally greater than 2 cm, and often more than 3 cm. The maps from the three processing centers are also more consistent in RL05 than RL04. GFZ_RL04 was generally noisier in the mid-latitudes than either CSR_RL04 or JPL_RL04, and JPL_RL04 had very large errors in the Atlantic Ocean (previously noted by Quinn and Ponte, 2010). To better quantify improvement we compute the variance reduction (Δvar) as

$$\Delta\mathrm{var} = 100 \times \frac{\mathrm{var}\left(\Delta_{\text{RL04-ECCO}}\right) - \mathrm{var}\left(\Delta_{\text{RL05-ECCO}}\right)}{\mathrm{var}\left(\Delta_{\text{RL04-ECCO}}\right)}, \tag{1}$$

where var($\Delta_{\text{RL04-ECCO}}$) is the variance of the residuals between RL04 maps and JPL_ECCO maps in each grid, var($\Delta_{\text{RL05-ECCO}}$) is the variance of the residuals between RL04 maps and JPL_ECCO maps in each grid, and the formulation computes the change relative to the variance in the old RL04 residuals as a percentage. If the variance in the RL05 residuals has become lower, then Δvar is positive,

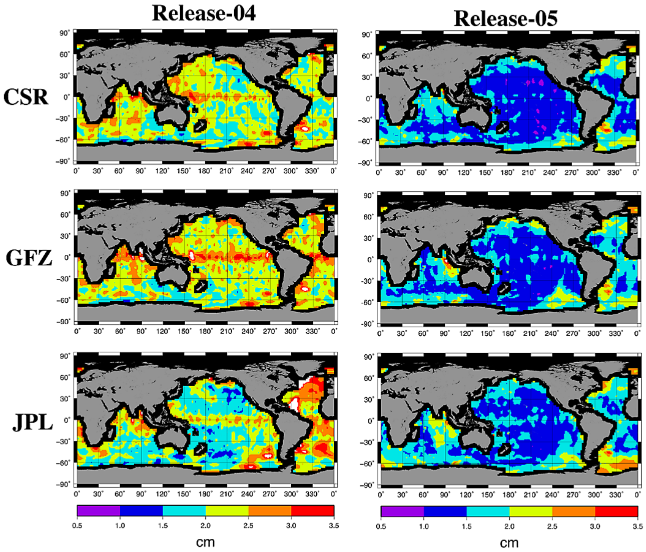

Fig. 3. Standard deviation of differences between unsmoothed OBP from JPL_ECCO and GRACE mapped OBP (destriped, 300 km Gaussian) for Release-04 (left column) and Release-05 (right column), using coefficients processed by CSR (top), GFZ (middle), and JPL (bottom).

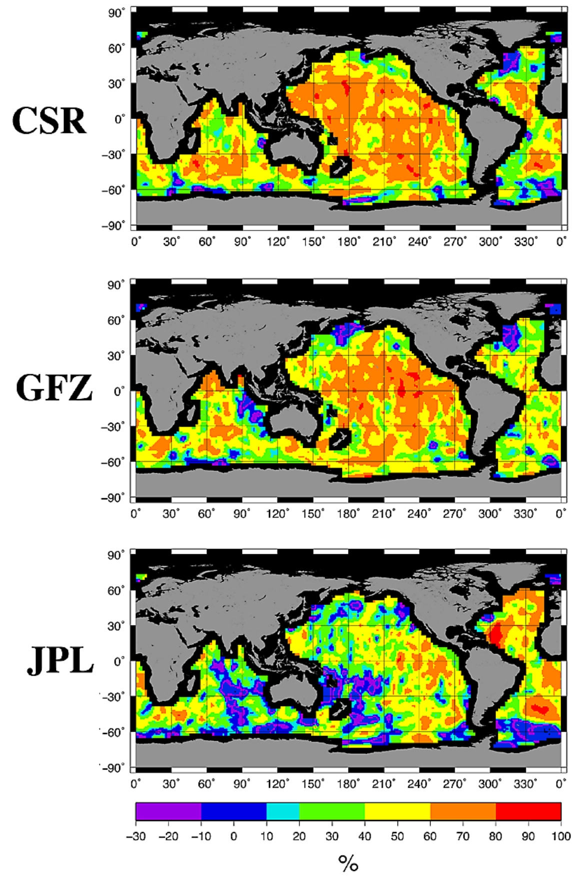

Fig. 4. Percent of variance reduced in Release-05 residuals compared to Release-04 residuals for coefficients processed by CSR (top), GFZ (middle), and JPL (bottom). Positive values mean the Release-05 residual variance is reduced, negative values mean that variance is increased relative to Release-04. All GRACE data were destriped and smoothed with a 300 km Gaussian. Please see text and Eq. (1) for details of the calculation.

indicating an improvement, while if it is negative, RL05 maps are more different from JPL_ECCO than RL04. Values are plotted in Fig. 4. The overall improvement, in terms of variance reduction relative to RL04 residuals, is between 50 % and 80 % over the majority of the ocean. The correlation between OBP from RL05 and that of JPL_ECCO is also significantly higher, with most values above 0.7 and many above 0.8 (Fig. 5).

As the JPL_ECCO model does not include the Arctic, this area will be examined using in situ data from a pair of Arctic bottom pressure recorders (ABPR) deployed at the North Pole by the North Pole Environmental Observatory program (Morison et al, 2007; data available from http://psc.apl.washington.edu/northpole/Data.html). Recorder ABPR1 (location: 89°15.26′ N, 60°21.58′ E) reported continually from 2005–2010. ABPR3 (location: 89°14.85′ N, 148°7.54′ E) reported continually from 2005–2008. The data were averaged over the first three years, detrended, and de-tided as explained by Peralta-Ferriz et al. (2011). The ABPR series is averaged into monthly points, to match the GRACE time resolution.

Average OBP from RL04 and RL05 GRACE data from CSR, JPL, and GFZ were computed in a 5° cap around the North Pole and compared to the ABPR data (Fig. 6). Correlations of the APBR data with GRACE are generally high. The correlation of APBR with CSR_RL04 was 0.86, which altered only slightly (to 0.89) upon updating to RL05. JPL's correlation with the APBR improved from 0.48 (RL04) to 0.87 (RL05). However, the correlation between the GFZ arctic data and the APBR decreased slightly from 0.87 for RL04 to 0.77 for RL05. Additionally, the variability of the GFZ_RL05 Arctic signal (2.0 cm) is only 70 % of the size of the other two RL05 GRACE signals (2.8 cm for CSR, 2.9 cm for JPL) and 58 % of the size of the ABPR variability (3.5 cm). The standard deviation of the residuals with the ABPR are 1.6 cm (CSR_RL05), 1.7 cm (JPL_RL05), and 2.3 cm (GFZ_RL05). These results suggest an unexplained reduction in real OBP variability in the GFZ_RL05 data that is not seen in either the CSR or JPL solutions. This is surprising, considering that the GFZ results are consistent with JPL and CSR in other basins (e.g., Fig. 4). One major difference between the CSR, JPL, and GFZ RL05 processing is the use of background ocean tide models. CSR and JPL use the GOT4.8 model, while GFZ uses the EOT11a version. This may be the source of the difference, and should be investigated further by the processing centers.

Because the results using RL05 coefficients from CSR, GFZ, and JPL are statistically identical in all areas but the

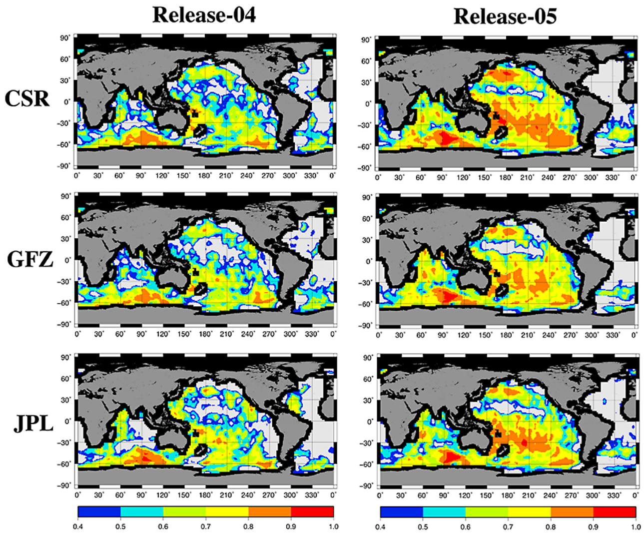

Fig. 5. Correlation between unsmoothed OBP from JPL_ECCO and GRACE mapped OBP (destriped, 300 km Gaussian) for Release-04 (left column) and Release-05 (right column), using coefficients processed by CSR (top), GFZ (middle), and JPL (bottom). Only values greater than 0.4 (the 99 % significance level) are shown.

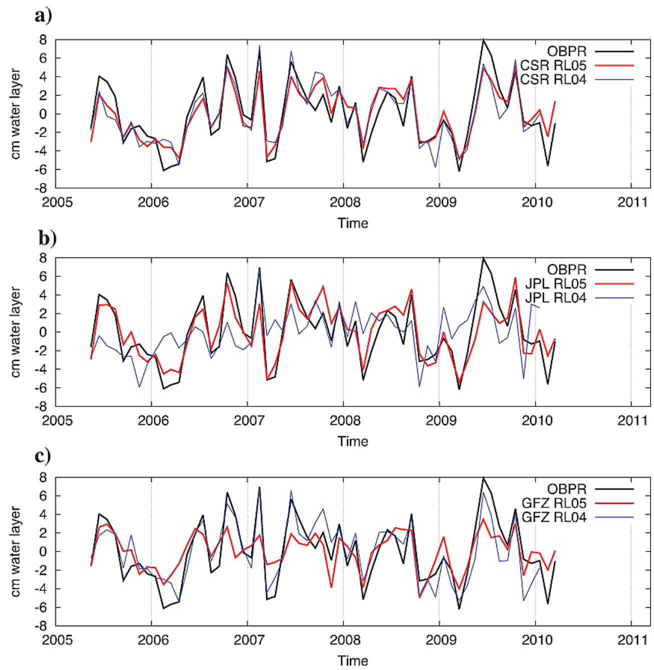

Fig. 6. Time series of OBP at the North Pole measured by a BPR (black line) and **(a)** CSR_RL04 & CSR_RL05, **(b)** JPL_RL04 & JPL_RL05, and **(c)** GFZ_RL04 & GFZ_RL05.

Arctic, for the remainder of this study, we will utilize only the CSR_RL05 grids to assess uncertainty and optimal smoothing. Although we do not show it, the results using the GFZ and JPL coefficients for the following tests are essentially the same, so our conclusions will apply to the new data from either CSR, GFZ, or JPL.

The RL04 maps on the GRACE *Tellus* website are currently produced using three different Gaussian smoothers in addition to the destriping algorithm (300 km, 500 km, 750 km). An analysis by Ponte et al. (2007) concluded that the 750 km version had lowest residuals with another version of the ECCO model. If we compare the RL05 data smoothed with different Gaussian smoothers to the unsmoothed JPL_ECCO OBP, we find that 500 km smoothed maps have significantly lower standard deviations (Fig. 7), compared to either the 300 km (shown in Fig. 3) or 750 km smoothing. The 300 km smoothing is still likely noisier, as evidenced by higher residuals in the tropics where the signal is low, while the 750 km signal is likely a more attenuating signal, as evidenced by an increase in residuals in the high-latitudes where OBP variability is highest in both the RL05 residuals (Fig. 7), and in residuals of the JPL_ECCO model smoothed with various Gaussians (Fig. 1). However, the 750 km smoother does noticeably reduce noise in the tropics where the signal is lower. We have examined additional smoothing radii between 300 km and 600 km, and 500 km does have a near minimum mean standard deviation of residuals of 1.3 cm, compared to 1.6 cm for 300 km. Note that some sort of destriping algorithm is still required; simply smoothing the coefficients with a 500 km Gaussian results in residuals with a mean standard deviation of 1.7 cm and differences that are larger than destriping and smoothing with a 300 km Gaussian (Fig. 7).

We can also compare the RL05 maps with the new de-aliasing model (Fig. 8). Note the significantly larger residuals at the higher latitudes in the Southern Ocean than when compared to JPL_ECCO. This is the area where the older de-aliasing model was shown to be deficient (e.g., Bonin and Chambers, 2011). Apparently, the new OMCT model still has issues in this region, but it appears that the GRACE data have corrected it so that they agree better with JPL_ECCO. JPL_ECCO likely performs better here because it assimilates altimetry, and these regions have strong barotropic signals that are reflected in sea level.

If we assume uncorrelated error between the GRACE data and ocean models and between the models, we can use the standard deviation of the residuals (σ) computed between the three different mapped data sets (JPL_ECCO, AOD, and GRACE) to estimate standard error (ε) in each set of data:

$$
\begin{aligned}
\sigma^2_{\text{E-A}} &= \varepsilon^2_{\text{E}} + \varepsilon^2_{\text{A}} \\
\sigma^2_{\text{G-E}} &= \varepsilon^2_{\text{G}} + \varepsilon^2_{\text{E}} \\
\sigma^2_{\text{G-A}} &= \varepsilon^2_{\text{G}} + \varepsilon^2_{\text{A}} \\
\varepsilon_{\text{G}} &= \sqrt{\frac{\sigma^2_{\text{G-A}} + \sigma^2_{\text{G-E}} - \sigma^2_{\text{E-A}}}{2}}\,, \\
\varepsilon_{\text{E}} &= \sqrt{\sigma^2_{\text{G-E}} - \varepsilon^2_{\text{G}}} \\
\varepsilon_{\text{A}} &= \sqrt{\sigma^2_{\text{G-A}} - \varepsilon^2_{\text{G}}}
\end{aligned}
\qquad (2)
$$

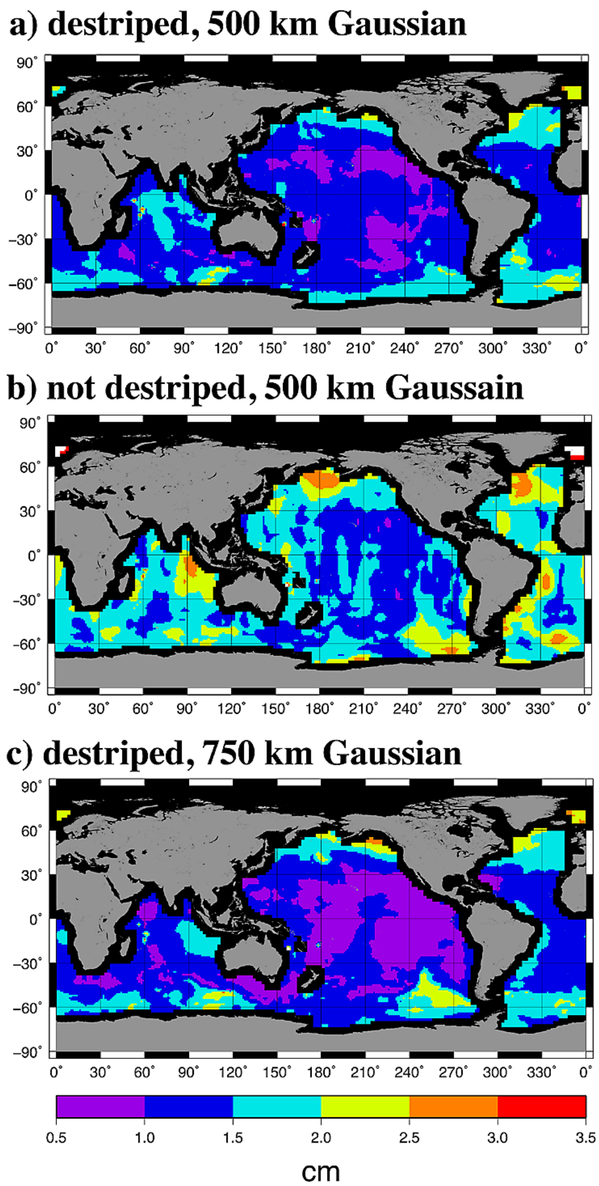

Fig. 7. Standard deviation of differences between unsmoothed OBP from JPL_ECCO and CSR_RL05 mapped OBP for **(a)** destriping and 500 km Gaussian, **(b)** no destriping and 500 km Gaussian, and **(c)** destriped and 750 km Gaussian.

where E, A, and G represent JPL_ECCO, AOD (i.e., OBP from GAD files), and GRACE, respectively. The two models share some common heritage, in the form of the starting primitive equations, and so will have some common errors. This means that the assumptions underlying Eq. (2) are not strictly valid. However, there are substantial differences in the models, such as the fact that JPL_ECCO assimilates data while AOD is simply a forced run, the winds, heat and freshwater fluxes come from two very different numerical weather prediction models, and parameterization of smaller-scale features (eddies, bathymetry) are different. Thus, we believe that

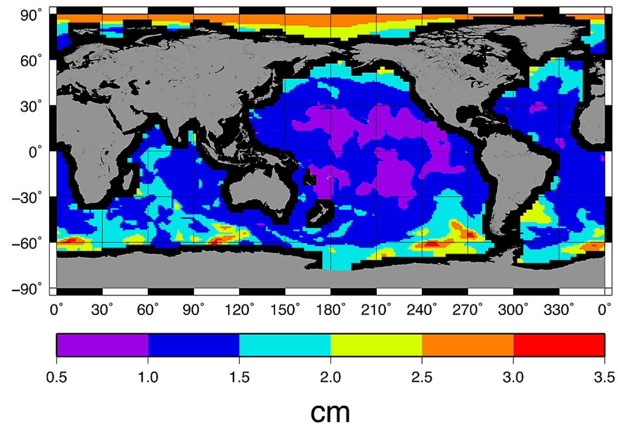

Fig. 8. Standard deviation of differences between unsmoothed OBP from RL05 AOD model and CSR_RL05 mapped OBP that has been destriped and smoothed with a 500 km Gaussian.

the differences are far larger than the potential common errors, and that Eq. (2) is a reasonable approximation to computing a better standard error in the GRACE maps then simply using the difference between GRACE and any one model, since this assumes no error in the model. The largest differences between the models are found in shallow waters and at high latitudes, while the smallest differences are found in the tropics where there are no significant OBP variations. The small difference in the tropics means that the computation of either ε_A or ε_E sometimes results in a negative sign. When this happens, we assume uncertainty is $\sigma_{\text{E-A}}$ (the difference between the models), which is an upper bound of the uncertainty, as it assumes one model has no error. The computation for ε_G does not suffer from this problem, but when the value of ε_A or ε_E is replaced, we also replace the value of ε_G computed in Eq. (2) with $\sigma_{\text{G-E}}$ (the difference between GRACE and JPL_ECCO), which again represents an upper bound of the error. This occurs in less than 5 % of the grids, all in the tropics.

Figure 9 shows the results of the calculation for the CSR_RL05 data, JPL_ECCO, and the monthly-averaged AOD data. The estimated standard error for GRACE is of order 1 cm over most of the mid-latitudes, which is significantly lower than previous releases of GRACE data, but still higher than the estimated uncertainty of the models. The uncertainty at high latitudes where OBP variability is large, especially in the Southern Ocean, is approximately the same on the GRACE maps as it is estimated to be in the JPL_ECCO model. Both are significantly smaller than the estimated uncertainty in the AOD model in regions of high OBP variability. We note that this applies only for the monthly and longer periods, and the same problem may not be seen in the AOD model at periods shorter than a month. This would require additional testing, which is beyond the scope of this discussion. GRACE still has high uncertainty in regions bordering ice sheets and glaciers (e.g., Greenland, Alaska, West

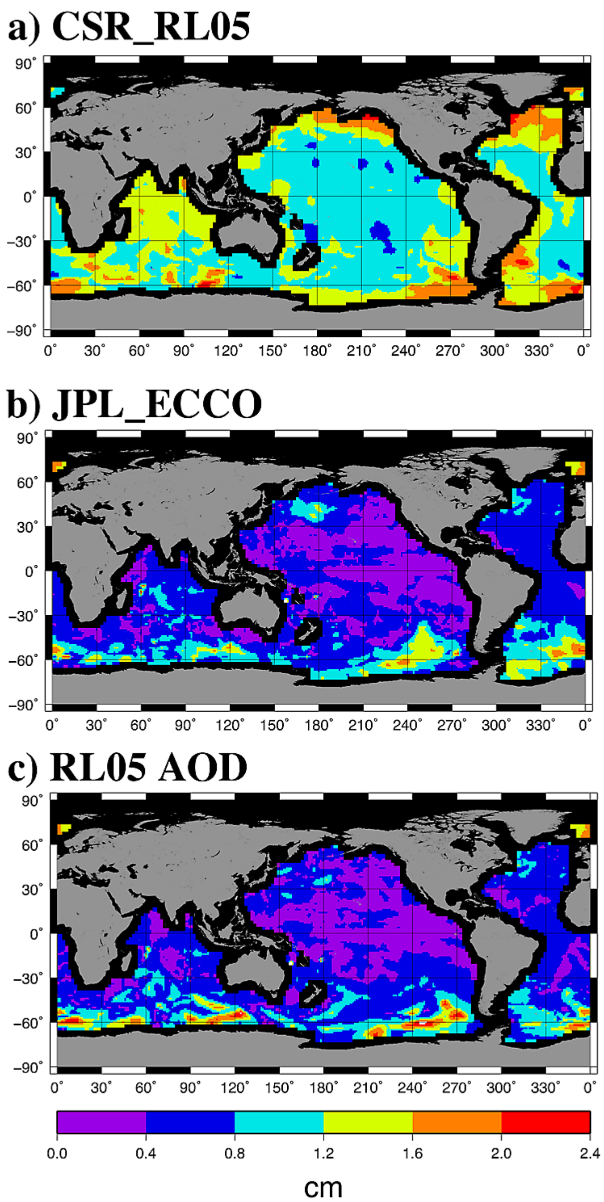

Fig. 9. Estimated standard error in mapped OBP based on Eq. (1) for **(a)** CSR_RL05 (destriped, 500 km Gaussian), **(b)** JPL_ECCO (unsmoothed), and **(c)** RL05 AOD (unsmoothed).

Antarctica), likely due to leakage of the large mass loss trends into the ocean. GRACE OBP in these areas should be used with caution.

We have also evaluated the GRACE OBP maps after projecting onto EOFs from a model, which has been used previously to reduce noise even further (Chambers and Willis, 2010). We tested using EOFs from both the new RL05 AOD model and JPL_ECCO, and ranging from using 10 EOF modes to 20 EOF modes. We found a minimum uncertainty estimate using 15 EOFs and patterns from the AOD model. This suggests that although the magnitude and variability of the AOD model OBP on monthly scales may not be as consistent with GRACE as JPL_ECCO is, the patterns of where the variability occurs is. Moreover, using the AOD model allows for the recovery of Arctic Ocean variability. The uncertainty of the EOF reconstructed (EOFR) OBP maps from GRACE is significantly lower than using the destriping algorithm alone, with a mean of 0.7 cm (Fig. 10). Uncertainty was computed using Eq. (2); the uncertainty estimated for JPL_ECCO and AOD is not shown as it is almost identical to that computed with the destriping algorithm (Fig. 9). Using the EOFR filtering reduces the error around ice sheets and glaciers dramatically, and also reduces noise in the mid-latitudes where OBP variability is low.

4 Tests of destriping algorithm

Now that the Release-05 data have been shown to be more accurate than Release-04 data using the same destriping algorithm, we test whether changing parameters of the algorithm will further reduce the uncertainty. For these tests, we will always truncate coefficients to degree/order 40 and use an additional 500 km smoother, as these have been shown to not significantly attenuate expected OBP variability in the deep ocean (Fig. 1), and give the lowest residuals with the current algorithm.

This leaves two parameters to adjust: the choice of lower-degree and order coefficients to leave unmodified, and the order of the polynomial. We compute a variety of destriped CSR_RL05 series, varying the onset of destriping from a minimum degree/order of 10 to 19, and the polynomial order from 2 (quadratic) to 7. We then compute the correlation of each 6-yr destriped set of GRACE maps with the unsmoothed JPL_ECCO maps, and also compute the standard deviation of the residuals. To determine which combination of destriping parameters best reduces differences between GRACE and JPL_ECCO, we examine the average correlation and standard deviation (Fig. 11). For comparison, with no destriping, but only a 500 km Gaussian smoothing applied, the average correlation between CSR_RL05 (to maximum degree/order 40) and JPL_ECOO is 0.67 and the standard deviation of the residuals is 1.7 cm.

The range of correlation and standard deviation for different parameterizations is relatively small, but an option with maximum correlation and minimum standard deviation can be found (Fig. 11). The old destriping used for RL04 utilized a fifth-order polynomial and a minimum filtered degree/order of 12, but the optimal parameterization for RL05 based on these tests is to start filtering at degree/order 15, and use a fourth-order polynomial for the fit (average correlation with JPL_ECCO: 70.7 %). The change to the residuals is small using the new RL05 filter; on average it reduces the variance by 10 %, although this can be as high as 50 % in some areas, notably in the Southern Ocean and near coastal North America (Fig. 12). Although the variance is increased in some areas,

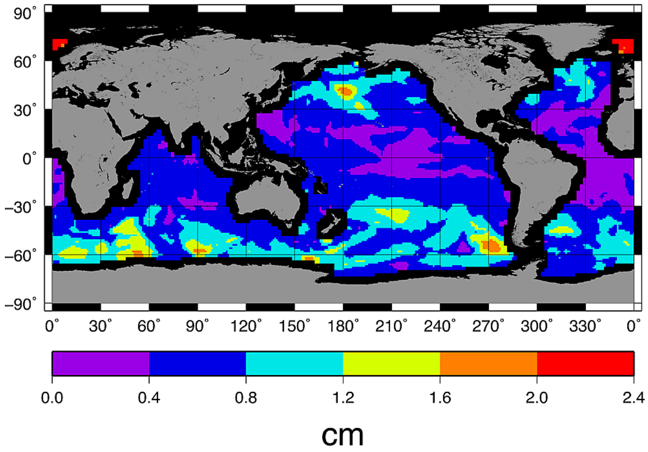

Fig. 10. Estimated standard error in mapped OBP based on Eq. (1) for EOFR filtered CSR_RL05 data.

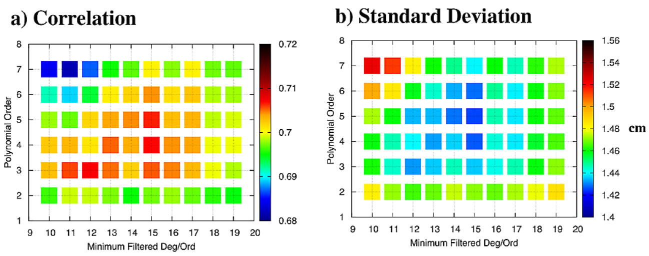

Fig. 11. Statistics comparing different destriping parameterizations with JPL_ECCO: **(a)** correlation, and **(b)** standard deviation of residuals, both averaged over ocean grids.

this is generally in areas where the variance is already low (standard deviation < 1 cm), so a 30 % increase in variance is less than 0.2 cm. We consider this acceptable since the filter reduces variance in many areas where the original variance was high (standard deviation > 2 cm).

5 Conclusions

The Release-05 processing is a significant step forward in reducing noise in the GRACE gravity coefficients. For the wavelengths that are most useful for studying ocean bottom pressure variations in the deep ocean (> 1000 km), an optimal destriping filter plus additional 500 km Gaussian smoother results in OBP that has an estimated standard error of ~ 1 cm over the mid- and low-latitudes, and between 1.5 and 2 cm at high-latitudes where OBP variations are high and around ice sheets and ocean-terminating glaciers. The uncertainty at high latitudes is slightly higher than that estimated for JPL_ECCO, but is less than that estimated for the atmosphere–ocean de-aliasing model used in GRACE processing. Applying a further filter by projecting the data onto EOF modes from a model reduces the uncertainty to a point where it is comparable to that estimated for JPL_ECCO.

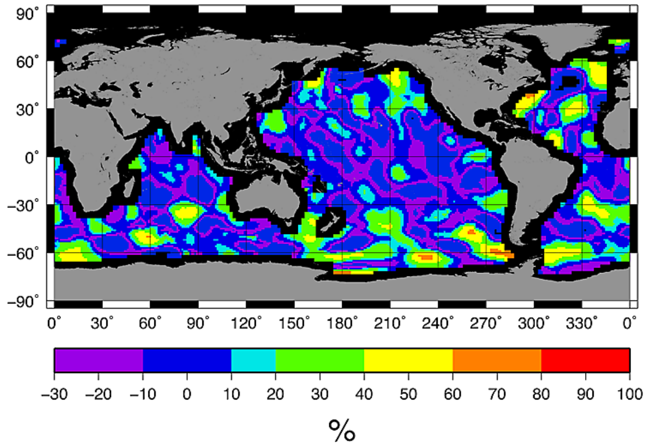

Fig. 12. Percent of variance reduced using new optimal destriping parameters and 500 km smoothing compared to those used for RL04, also using 500 km Gaussian smoothing. Positive values mean the variance with the new algorithm is reduced, negative values mean that variance is increased.

Results are virtually the same for data from all three processing centers (CSR, GFZ, JPL) except in the Arctic, where there is evidence that GFZ has lower signal than expected, but CSR and JPL have similar variability as a bottom pressure recorder. Although the modified destriping filter that is proposed does increase variance of residuals with JPL_ECCO in some areas, the overall average reduction is positive (especially in the Southern Ocean), and we believe it is better to under-filter the GRACE data than to over-filter it.

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