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Regional distribution of steric and mass contributions to sea level changes

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ABSTRACT

The contributing factors to regional sea level variability have been explored for the period 2004–2008 based on altimetry observations, hydrographic data and GRACE measurements. The regional averaged annual cycle of the mass contribution to sea level is shown to be highly unsteady. When compared with steric-corrected altimetry, both signals are coherent, though in some regions the coherence analysis is limited by the use of interpolated hydrographic data and in the equatorial regions it is limited by the low signal-to-noise ratio of GRACE data. The closure of regional sea level budgets depends mainly on the GIA correction chosen. A reconstructed global sea level field (with the atmospheric signal eliminated) spanning the second half of the 20th century together with historical hydrographic observations are used to infer the regional mass contribution to sea level rise for the last decades. Results indicate that mass addition from continental ice is the major contributor to regional mean sea level rise for the last decades. In addition, the spatial patterns of mass rates of change point at Greenland as the main source of fresh water input.

1. Introduction

Long-term global sea level changes have been routinely estimated based on tide gauge measurements with a biased spatial distribution. Since 1992 satellite altimetry has revealed a high spatial heterogeneity of sea level changes, with areas experiencing sea level rise up to three times larger than the global rate and others where sea level has dropped (Cazenave and Nerem, 2004; Cazenave and Llovel, 2010). The contributors to long-term global sea level changes are steric changes and mass addition/subtraction. At regional scale we must add the mass displacements due to the mechanical atmospheric forcing and changes in the oceanic circulation, which may play a significant role in those areas where circulation features are important and determine local sea level (e.g., Gulf Stream or Kuroshio). Separating the sources and reducing the uncertainties in the quantification of each contributor to regional sea level changes is of key importance to understand the causes of sea level variations and to infer future changes.

The global sea level budget has been explored by different authors. Willis et al. (2008) did it on the basis of altimetry, in-situ hydrographic data from Argo floats and space gravimetry observations from the Gravity Recovery and Climate Experiment (GRACE) mission between 2003 and 2007. They concluded that while intra- and inter-annual changes inferred from the different observation sets are consistent, the trends computed for the analyzed period do not agree. Conversely, Leuliette and Miller (2009) using the same data for a slightly different period (2004–2007) found statistical agreement between observed sea level rates of change and the addition of the steric and mass components. Cazenave et al. (2009) also found consistency between steric sea level as inferred subtracting GRACE from altimetry and as observed from Argo floats for the period 2004–2008, respectively. Cazenave et al. (2009) pointed to the critical contribution of the Glacial Isostatic Adjustment (GIA) correction that has to be applied to raw GRACE data as one of the reasons for the disagreement between different authors. The GIA correction is based on solid earth models with a particular rheological profile, ice history deglaciation chronology of the late-Pleistocene ice sheets and defined parameters of the visco-elastic properties of the Earth. GIA reflects in the GRACE signal as a long term trend in the gravity field that is not due to the instantaneous redistribution of water over the Earth's surface. It is thus necessary to separate that trend from actual changes in the water content. This linear correction determines to a large extent the rates of change of the ocean mass component inferred from GRACE data. There are currently two broadly used solutions available for such correction, Paulson et al. (2007) and Peltier (2004) models, with very different global rates (1 and 2 mm/yr, respectively) and even larger differences at regional scale. The differences between the two models are analyzed in Peltier (2009), Peltier and Luthcke (2009) and more recently in Chambers et al. (2010). Chambers et al. (2010) have found that the differences are mostly attributed to large trends in predicted degree-2, order-1 geoid coefficients in the Peltier (2009) model. Peltier and Luthcke (2009) attributed these large rates to present-day ice losses. However Chambers et al. (2010) showed that the signals in Peltier's model are inconsistent with the polar motion and rotation feedback theory he claims to be using and considered that these rates are unrealistic.

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Also the use of different time-period and processing techniques are partly responsible for the differences found between different studies. For instance, Leuliette and Miller (2009) showed that results were sensitive on how the Argo data are mapped in the early part of the record. This is further exemplified by the fact that, even over the same time-period, Leuliette and Miller (2009) and Cazenave et al. (2009) Argo results differed by 0.5 mm/year.

In this paper we address the quantification of the contributions driving sea level variability regionally rather than globally. When coastal protection and impact assessment are concerned, regional sea level rates of change are of key importance to understand and project how sea level changes will affect a particular area. Given the high spatial heterogeneity of sea level variability the global rates become meaningless in this context. The closure of the regional sea level budget has been explored by fewer authors. Llovel et al. (2010) found a poor agreement between the regional patterns of steric-corrected altimetry and those of the mass contribution inferred from GRACE. Conversely, in the Mediterranean Sea observations of total sea level and its components are reported to be consistent (Fenoglio-Marc et al., 2006; Calafat et al., 2010). In a further step, Calafat et al. (2010) took advantage of this good agreement to infer the mass contribution to Mediterranean mean sea level for the second half of the 20th century, using a reduced-space optimal interpolation of altimetry and tide gauge data to infer total sea level fields for the pre-altimetric period.

In this work we extend the methodology applied in the Mediterranean Sea by Calafat et al. (2010) to different regions worldwide. The independent measures of the mass contribution to sea level provided by GRACE since 2002 are combined with estimates of steric sea level and observed total sea level to explore the regional sea level budgets. The goal of this study is to investigate to which extent regionally averaged mass variations are mimicked by steric-corrected altimetry in different areas and at different temporal scales. Additionally, in order to infer the mass contribution during the last decades of the 20th century we use a global reconstruction of sea level fields together with historical hydrographic observations. Given the assumptions inherent to this methodology a careful uncertainty assessment is carried out.

2. Data sets

2.1. Sea level

Gridded monthly sea level anomalies with a map spacing of 1°×1° computed from satellite multimission with respect to a seven year mean were obtained from the AVISO data server (http://www.aviso.oceanobs.com). This data set spans the period from October 1992 to present. All geophysical corrections have been applied; the atmospheric correction is applied using the Dynamic Atmospheric Correction currently delivered by AVISO, which consists of using the barotropic model MOG2D (Carrère and Lyard, 2003) to correct frequencies greater than 20 days and the inverted barometer approach otherwise.

The sea level anomaly fields obtained from altimetry data have also been combined with tide gauge records (from which the atmospheric signal has been previously removed) to obtain reconstructed global sea level fields for the period 1950–2003. This field was computed by combining selected 99 tide gauge records (from Llovel et al., 2009) and 11 years of altimetry observations over 1993–2003. The method is based on the reduced space optimal interpolation described by Kaplan et al. (1998, 2000). It uses the spatial structure (EOFs) of the sea level field obtained from the 2-D well resolved spatial fields of altimetry satellite measurement to interpolate the historical measurements from tide gauge records. Following Church et al. (2004), in this run we added a spatially uniform EOF (referred to as EOF0) to the set of EOFs computed from altimetry; the inclusion of this extra EOF is aimed to account for any basin-uniform movement. The global reconstructed sea level fields are mapped on a (1°×1°) over the period 1950–2003.

2.2. Steric sea level

Steric sea level fields were computed using two data sets: the Ishii global gridded temperature (T) and salinity (S) climatology (Ishii and Kimoto, 2009) and the ENACT/ENSEMBLES version 2a (EN3) data set (Ingleby and Huddleston, 2007) made available by the Met Office Hadley Centre. The Ishii data set has been produced by objective analysis of in-situ observations and consists of monthly gridded T, S fields with a spacing of 1°×1°; the vertical domain extends down to 700 m, with data on 16 levels. This data base covers the whole second half of the 20th century, namely the period 1945–2006. The steric sea level component has been computed at each grid point by integrating the specific volume down to 700 m. The Ishii climatology includes an estimation of the uncertainties associated with the interpolated T and S gridded fields, which can be propagated to obtain the uncertainty of the steric component.

The EN3 data set has been produced by objective analysis of the T and S profiles of the World Ocean Database'05, the Global Temperature and Salinity Profile Project, Argo and the Arctic Synoptic Basin-Wide Oceanography Project. In the current version (v2a) the Argo profiles with erroneous pressure values according to Willis et al. (2009) and profiles that are suspect of containing errors according to Guinehut et al. (2009) have been rejected. The part of the database used in this work consists on monthly gridded T, S fields with a map spacing of 1°×1°; the vertical domain extends down to 970 m, with data on 24 levels. The time period spanned by this data base is 2002–2008, i.e., it extends to years further than Ishii, which is important when considering the short period spanned by GRACE data.

The computation of the steric component is thus restricted to the upper part of the ocean: 700 m in the case of the Ishii data set and the top 970 m for EN3. The reason why the data bases do not include deeper fields is that below those depths the number of observations decreases significantly, making the interpolation unreliable (Leuliette and Miller, 2009; Dhomps et al., 2010). Dhomps et al. (2010) reported that integrating steric sea level down to 1000 m recovers at least 80% of the total signal worldwide.

2.3. GRACE measurements

Finally the mass contribution has been obtained from measurements provided by the GRACE mission launched in 2002. GRACE measures the variations in the gravity field caused by changes in the water mass of the Earth, then providing an independent measure of the mass contribution to sea level changes. The Level-2 Release-04 (RL04) gravity coefficients computed at the Center for Space Research (CSR) were used to estimate monthly global water mass variations for the period August 2002 to the end of 2008 with a spatial grid of 1°×1°. The data include corrections to specific spherical harmonic coefficients due to solid Earth and ocean tidal contributions to the geopotential. GRACE pre-processing also removes variability from an ocean barotropic model (i.e., the high-frequency ocean mass variations forced by winds and pressure) along with the atmospheric mass. The solid and ocean pole tide are also removed. RL04 coefficients are supplied to degree and order 60. Correlated errors between even or odd Stokes coefficients (C_{lm},S_{lm}) are removed by means of a 5th order polynomial fit (Chambers, 2006). Degree 2, order 0 coefficients from GRACE are replaced with those from the analysis of Satellite Laser Ranging (SLR) data (Cheng and Tapley, 2004). We also restore modeled rates for certain coefficients (degrees 2, 3, and 4 for order 0, and degree 2 for order 1) as discussed in the Processing Standards Documents (Bettadpur, 2007). The last step for obtaining the Stokes coefficients is done by adding back the mean monthly gravity coefficients of the ocean bottom pressure supplied by the project which were removed in the preprocessing (Flechtner, 2007) and an estimate of degree 1 gravity coefficients (Swenson et al., 2008).

In order to compensate for poorly known short-wavelength spherical harmonic coefficients, gravity coefficients are converted into smoothed maps of surface mass density by means of a Gaussian spatial average (Wahr et al., 1998). Surface mass density is converted to equivalent surface height by dividing it by the density of fresh water. The radius of the Gaussian smoothing function used in this study is 500 km. Because the smoothing is done on global spherical harmonics, any large hydrological signal over land will leak into the ocean signal near the coast. The Climate Prediction Center (CPC) hydrological model has been used to correct the effect of land waters (Fan and van den Dool, 2004). In order to be consistent with GRACE, we have smoothed the hydrology field by using the same spatial averaging applied to GRACE data.

A correction for the Glacial Isostatic Adjustment (GIA) has also been applied. In this work we use the GIA correction field computed by Paulson et al. (2007) and expressed in terms of a mass rate. It is the only solution which is currently publicly available and it has been obtained from http://grace.jpl.nasa.gov/data/pgr. Again, the mass rate estimates were smoothed using a Gaussian averaging function of 500 km radius in order to be consistent with GRACE data.

3. Regional sea level variability during 2002–2008

Sea level anomalies from altimetry and the steric and mass contributions have been compared for the period 2004–2008. Such period has been chosen to ensure that the number of S observations is large enough to compute steric sea level reliably. The Ishii data set is thus restricted to the short period 2004–2006. Despite this limitation it is included in the analysis because it provides an estimation of the uncertainties that will be used later on in the paper. For the purpose of comparison all fields have been filtered using a Gaussian filter of radius 500 km. The steric contribution has been estimated using the two available data sets, referred to simply as Ishii (2004–2006) and EN3 (2004–2008) hereinafter.

3.1. The seasonal cycle

The dominant signal in the time series is the seasonal cycle caused by the warming/cooling of the ocean and the exchange of waters between land and oceans between seasons. Since the sea level seasonal cycle is unsteady in time (Marcos and Tsimplis, 2007) the same period (from 2004 to 2008) has been chosen in all cases for the consistency of comparisons.

The annual cycle of the mass component has been obtained on one hand from GRACE observations and on the other hand from steric-corrected altimetry (Fig. 1), ranging between 0 and 6 cm in both cases (color scales are however defined from 0 to 4 cm for a better visualization of spatial structures). On average the annual cycle represents 25% of the monthly mass signal, reaching values of up to 50% only in the southern ocean, where amplitudes reach 5 cm. According to GRACE observations, maxima values are found around Greenland and in the northernmost Pacific coasts; none of these areas are monitored by the altimetry data used, therefore preventing the comparison with non-steric sea level. Leakage from land hydrology is expected to be very large in this areas and it may generate such signal. For the same reason, larger than average values are also found close to Antarctica, especially around the Antarctic Peninsula. An annual signal larger than average is also obtained in the western equatorial regions of the North Atlantic, likely related to the seasonal variations of large river runoff. Large differences are found between GRACE observations and steric-corrected altimetry at equatorial regions. We suspect this is

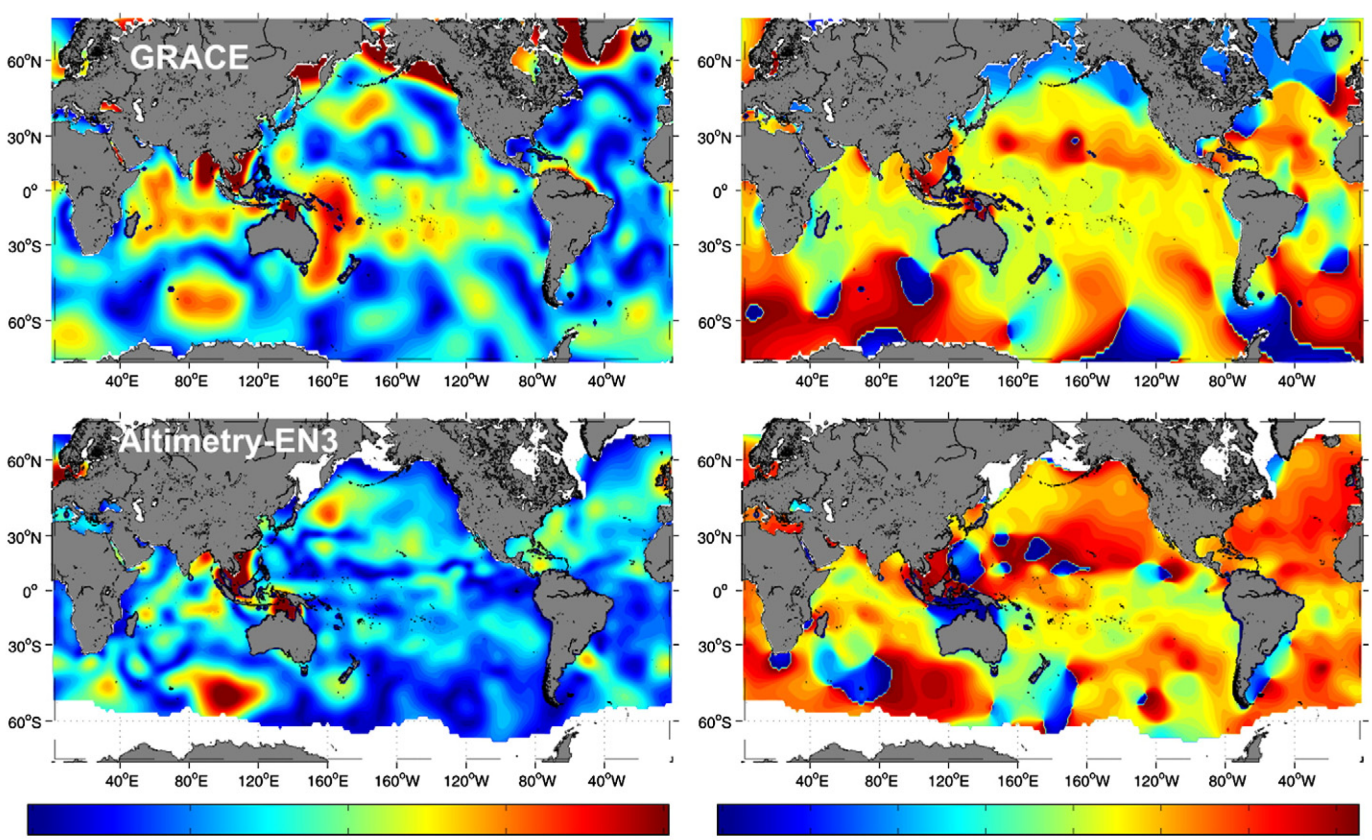

Fig. 1. Mean annual amplitudes (left column) and phases (right column) of the mass component of sea level as observed by GRACE and as inferred from altimetry minus steric sea level. Values are in cm and degrees respectively.

related to different spatial resolution of the data and to the relatively low signal-to-noise ratio of GRACE in the tropics (e.g., Wahr et al., 2004).

3.2. Regional sea level budgets

Regional sea level budgets are explored for seven regions, namely the North, Equatorial and South Pacific and Atlantic Oceans and the Indian Ocean. The mass contributions to sea level changes averaged over each region as observed by GRACE and as inferred from steric-corrected altimetry using Ishii and EN3 climatologies are shown in Fig. 2. The correlation between observed and computed mass for their common periods is quoted for each graph. Time series are dominated by seasonal variations. Correlations are significant almost everywhere with values ranging between 0.4 and 0.8. The exceptions are the north and equatorial Atlantic regions, where correlations are not statistically significant for the longer period 2004–2008. This could be partly attributed to the computation of steric sea level using interpolated data that are biased in such regions with high variability due to the Gulf Stream (Miller and Douglas, 2004). On the other hand in these regions there is a smaller number of valid S profiles which suggests a less reliable steric estimation.

Overall it is fair to say that the two steric sea level data sets are consistent to each other and provide similar correlations between steric-corrected altimetry and mass changes from GRACE.

A large part of the high correlation values is likely due to the dominance of the seasonal cycle, with a major portion of the seasonal cycle controlled by global ocean mass variations, of ± 1 cm. In order to explore the inter-annual consistency of sea level budgets, the mean seasonal cycle is removed from each time series. Results are shown in Fig. 3 (only the steric-corrected altimetry using EN3 is shown for simplicity). De-seasoned time series of mass observations and steric-corrected altimetry have similar variability. Variances are larger for steric-corrected altimetry than for GRACE data only in the south Atlantic (2.1 and 1.2 cm^2 respectively) and in the Indian Ocean (0.9 and 0.5 cm^2 respectively). They are nearly the same everywhere else, ranging between 0.3 and 0.9 cm^2 depending on the region. The RMS of the signals and the RMS of their difference is generally of the same magnitude. The reason is the presence of some large peaks in one of the signals and not in the other.

As expected, correlations have decreased in all the regions when not considering the seasonal cycle (Fig. 3). They become not significant in the equatorial regions and in the North Atlantic. Smoothed time series obtained with a 6-month running average are also plotted in Fig. 3. Linear trends from deseasoned time series are

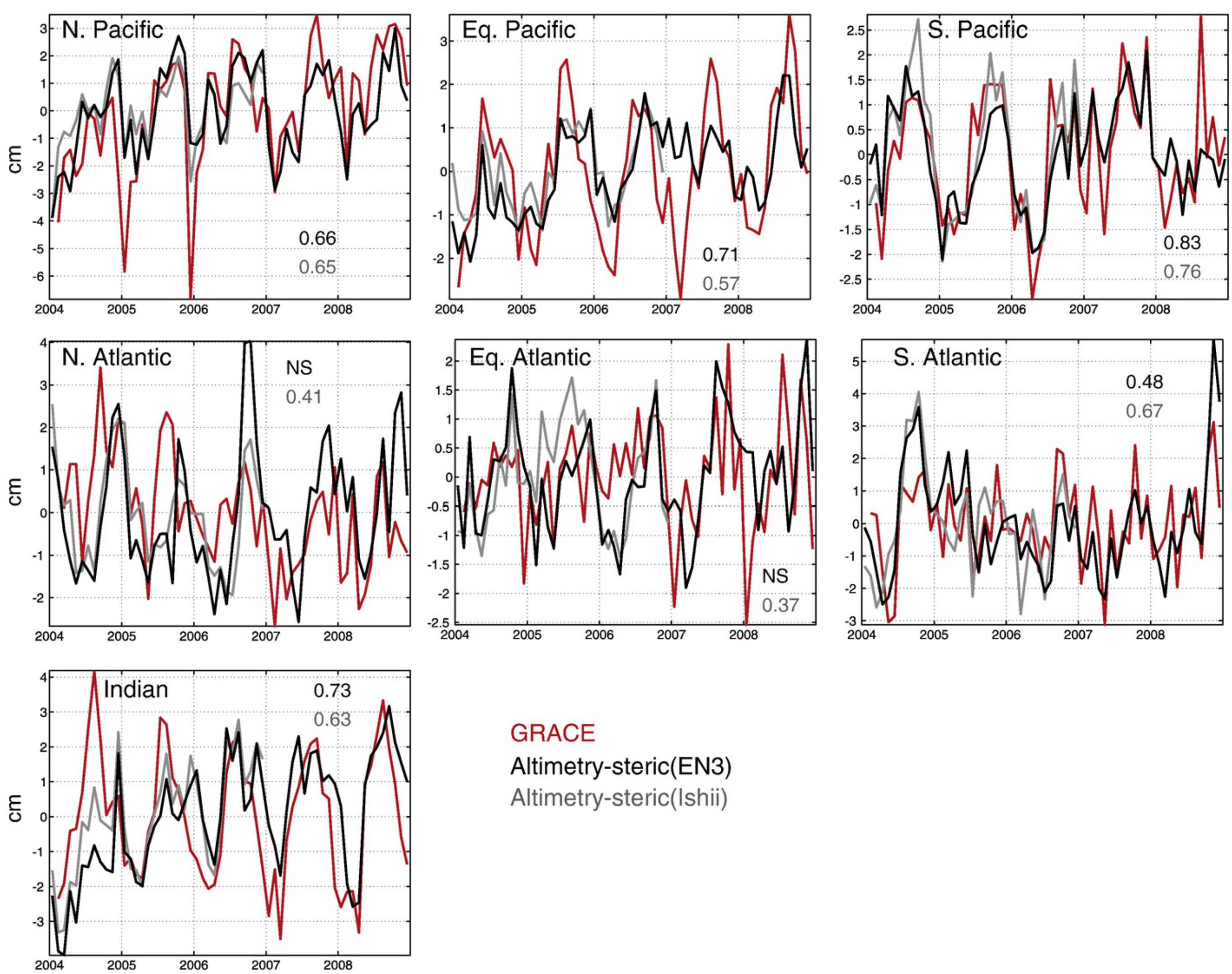

Fig. 2. Comparison between the measured (from GRACE) and inferred (altimetry minus steric component) mass contribution to sea level for different regions. North regions are defined between latitudes 30°N–50°N, equatorial between 20°S–20°N, and south 60°S–30°S. The correlation between both curves is quoted in the right low corner of each graph when significant at the 99% confidence level (NS otherwise).

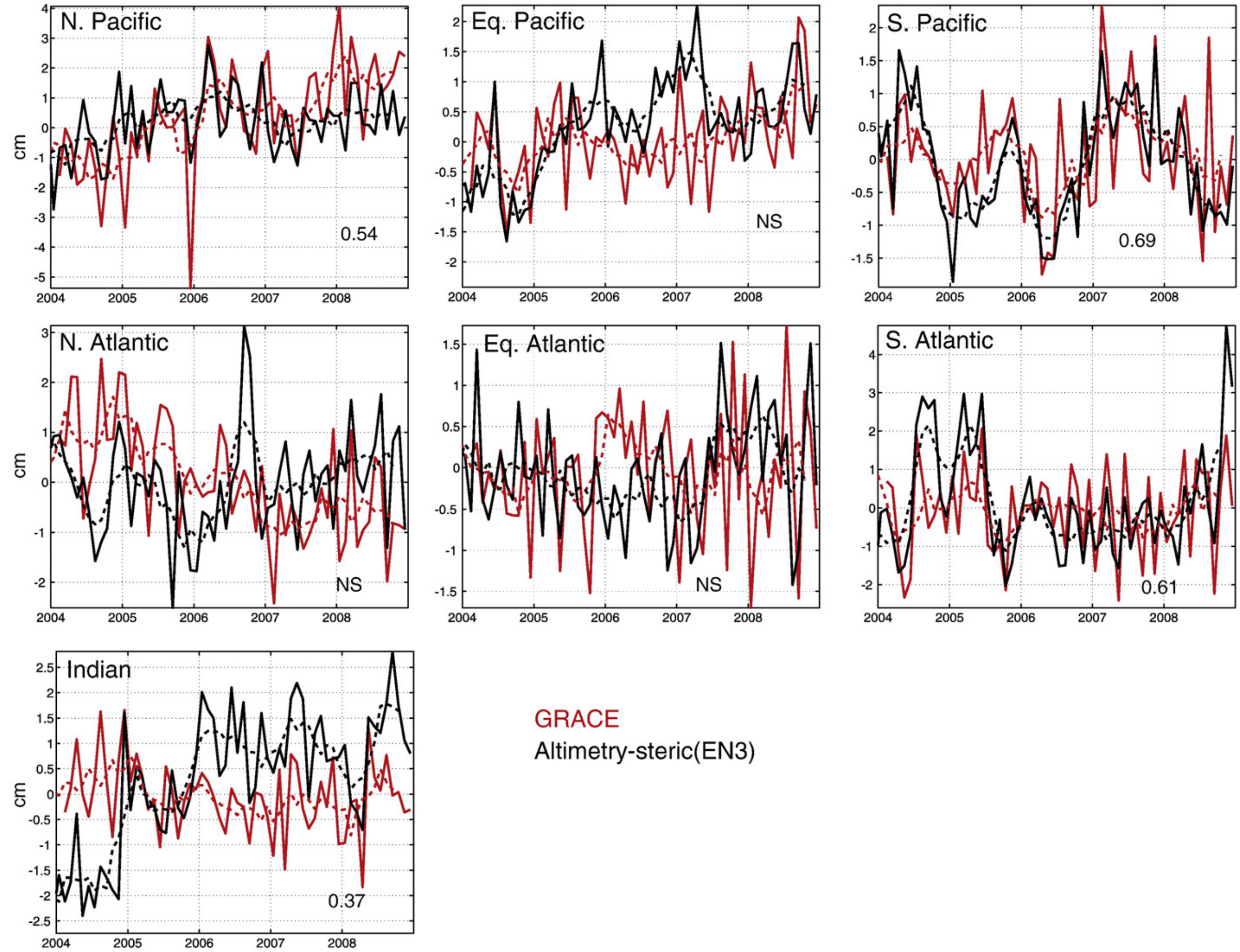

Fig. 3. Steric-corrected altimetry obtained using the EN3 data base (black lines) and mass contribution from GRACE (red lines) de-seasoned (solid) and filtered with a 6-months running mean (dashed). Correlation coefficients are quoted for each region.

quoted in Table 1 together with uncertainties as given by standard errors.

In the North Pacific region, inter-annual changes of steric-corrected altimetry and GRACE observations are fairly correlated (0.54) and both show positive linear trends. GRACE time series present a relative minimum in 2006, in agreement with the results of Chambers and Willis (2008) for a similar but smaller region of the North Pacific. These authors found a trend of about 9 mm/yr for a different period (2003–mid 2007). In our case the GRACE trend for 2004–2008 is 6.7 ± 1.1 mm/yr. Song and Zlotnicki (2008) suggested that ocean bottom pressure below the sub-polar gyre of the North Pacific correlates with tropical ENSO episodes, resulting in below average ocean bottom pressure shortly after an event and above average shortly before. Fig. 4a represents smoothed and detrended GRACE observations averaged over the North Pacific altogether with the multivariate ENSO index (Wolter and Timlin, 1998). Two strong ENSO events took place during the GRACE period, one in early 2003 and one in 2007 (see Fig. 4). Despite there is not statistically significant correlation between the two curves, GRACE observations are qualitatively consistent with Song and Zlotnicki (2008) hypothesis for these both events. This was already partly confirmed by Chambers and Willis (2008), but only until mid-2007. The longer GRACE time series used here permits confirming the predicted drop in ocean bottom pressure during 2007, though this does not discard the possibility that such changes can be due to inter-annual variations not related to ENSO episodes.

In the southern Pacific the correlation between observed and inferred mass variations at inter-annual scales reaches 0.7. No significant trends are found in any of the time series (Table 1). Additionally, Fig. 4b evidences the relationship between mass changes in the southern Pacific basin and the ENSO variability, with a correlation of −0.5 at a 6-months lag.

In the Indian Ocean GRACE observations and steric-corrected altimetry show a significant correlation of 0.4. However, large differences are found in their trends (Table 1). While GRACE observes

Table 1
Trends of the regionally-averaged mass contribution to sea level estimated from GRACE and from steric-corrected altimetry (using the EN3 climatology) for the period 2004–2008. Units are in mm/yr.

Region	GRACE	Steric-corrected altimetry
N. Pacific	6.71 ± 1.11	1.72 ± 0.95
Eq. Pacific	1.60 ± 0.66	3.36 ± 0.62
S. Pacific	0.20 ± 0.80	-0.18 ± 0.83
N. Atlantic	-4.91 ± 0.86	0.93 ± 0.87
Eq. Atlantic	0.09 ± 0.71	1.06 ± 0.58
S. Atlantic	0.54 ± 1.11	-1.29 ± 1.26
Indian	-0.80 ± 0.64	6.06 ± 0.95

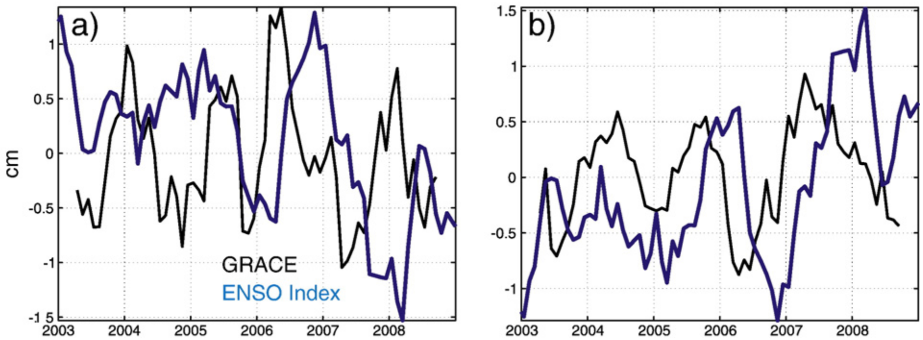

Fig. 4. Detrended and smoothed GRACE observations averaged over the North and South Pacific (black, in cm) and ENSO index (blue).

a trend only slightly different from zero, the value obtained from steric-corrected altimetry is much larger (6.1 ± 1.0 mm/yr). Such discrepancy was already pointed out by Willis et al. (2008). They noted that the large trend observed in altimetry was not visible in steric data, thus pointing at a mass exchange as being the main cause. However, this is not confirmed by GRACE observations. Recent investigations point at new pressure biases in the instruments deployed in the Indian Ocean as the origin of the difference (D. Chambers, personal communication).

Mass exchanges between Atlantic and Pacific regions are plotted in Fig. 5 in order to explore the sub-basin inter-annual variability. Only GRACE time series are used to avoid the unrealistic steric sea level estimates in the north Atlantic region. Chambers and Willis (2008) already demonstrated that inter-annual mass exchanges as large as seasonal variations exist between the Atlantic and Indian basins with the Pacific. Also Stepanov and Hughes (2006) identified mass exchanges between the Southern Ocean and the Pacific (northward 35°S). We therefore focus here in sub-basin exchanges between northern and southern latitudes. Fig. 5 reveals mass exchanges between the target regions. Interestingly, two different regimes of inter-annual barotropic oscillations can be identified. For the period 2003–2006 the Pacific Ocean oscillates in phase while the north and south Atlantic oscillate out of phase. From 2006 onwards the behavior is the opposite with the Pacific showing clear out of phase signals and the Atlantic oscillating coherently. Whether this shift is an exceptional event or not can only be determined with a longer time series not yet available. The reasons thus remain uncertain and clearly further research is needed to determine its origin.

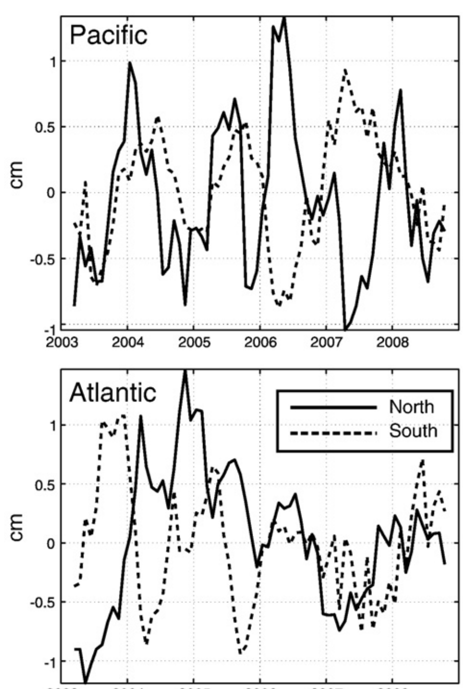

Fig. 5. Detrended and smoothed (with a 6-months running mean) averaged GRACE observations over the northern and southern sub-basins of the Pacific (top) and Atlantic (bottom) Oceans.

3.3. Consistency of inter-annual variations

Changes in steric-corrected altimetry and mass variations from GRACE at inter-annual scales are compared on the basis of regional EOFs. The reason why regional analysis has been preferred to global analysis is to account for basin scale mass changes and regional processes without being masked by large scale ocean variations. De-seasoned fields of altimetry and steric sea level from EN3 as well as mass variations from GRACE are used. All fields are filtered using a Gaussian filter of radius 500 km to be consistent with each other. EOFs have been computed for the same seven regions defined above. However, only results for the most interesting areas, namely north and south Pacific and Atlantic Oceans and for the Indian Ocean, are shown (Figs. 6 to 10).

The two leading EOFs of the North Pacific region are shown in Fig. 6. The first EOF explains significantly more variance in the GRACE decomposition (53%) than in the steric-corrected altimetry decomposition (30%), but the patterns are similar. Positive values dominate in the entire domain, being larger in the western area, coinciding with the region where Chambers and Willis (2008) found larger trends. The large trend found in GRACE data is entirely explained by the first EOF (8.4 ± 2.0 mm/yr) and is thus associated to the corresponding spatial pattern. In the second EOF a dipole structure is observed in steric-corrected altimetry, whereas GRACE field presents a track-like pattern and does not represent a physical signal.

In the south Pacific (Fig. 7) the first EOFs explain the same variances in steric-corrected altimetry than in GRACE (30%). A SE–NW gradient is found in the spatial patterns in both cases, although steric-corrected altimetry has larger values in the NW. The linear trends of the temporal amplitudes are large (6 and 11 mm/yr, respectively), despite the trend of the total series is not different from zero. The second EOF also shows similar patterns in the two fields and, in this case, also similar temporal amplitudes. Spatial patterns of mass variations reflect the signature of the El Niño, the dominant climatic mode in the area. Correlations of the first and second amplitudes of GRACE data present statistically significant correlations with ENSO index of 0.6.

The north Atlantic decomposition shows clear discrepancies between steric-corrected altimetry and GRACE spatial EOFs (Fig. 8). The main signal of the steric-corrected altimetry leading EOFs is associated with the Gulf Stream variability. This happens because the use of interpolated gridded data for estimating steric sea level biases the values with respect to using single T and S profiles (Miller and Douglas, 2004). Therefore the mass contribution in the North Atlantic as inferred from steric-corrected altimetry is not reliable. The temporal amplitude of the first EOF computed from GRACE observations is significantly correlated (0.50) with the East Atlantic pattern (Barnston and Livezey, 1987). This climate pattern is a dominant mode in the North Atlantic consisting in a NE–SW dipole similar to NAO. The pattern of the first EOF presents the same structure. Notably, the first mode is not correlated with the NAO index. We attribute it to the fact that NAO acts over northernmost latitudes.

In the south Atlantic the spatial patterns and the variances explained of the two leading EOFs are consistent between steric-corrected altimetry and GRACE (Fig. 9). The signals found were first identified by Fu et al. (2001) as a free barotropic mode with a length of about 1000 km and a period of 25 days and with strong seasonal and inter-annual variability, on the basis of altimetric measurements and theoretical considerations. Hughes et al. (2007) reported a mode with lower period (20 days) and suggested that its variability is due to interaction between eddies, mean flow and topography rather than to direct atmospheric forcing through pressure and wind. Weijer et al. (2007) found that the flow variability in the Argentine Basin is caused by the excitation of several barotropic normal modes of this basin. The presence of multiple oscillatory basin modes would reconcile the previous frequencies. Interestingly, the first EOF of steric-corrected altimetry clearly reproduces the dipole pattern found by Fu et al. (2001) and later on confirmed by Weijer et al. (2007).

In the Indian Ocean the largest feature of the GRACE decomposition is found in the north-eastern part of the domain and is related to the gravity variations generated by the Sumatra earthquake in 2004 (Fig. 10) (Han et al., 2010); it is thus not reproduced by the steric-corrected altimetry. The spatial patterns of the first EOF, accounting for nearly the same amount of variance for the two data sets, present in both cases larger values in the eastern part of the domain. However, structures in steric-corrected altimetry are smaller and do not appear in GRACE. The second EOF of the GRACE decomposition shows marked track-like structures.

In summary, at inter-annual scales the steric-corrected altimetry is consistent with observations of mass changes in the north and south Pacific and in the south Atlantic. Results are not conclusive for equatorial areas and are clearly non-consistent in the north Atlantic, especially near the Gulf Stream. In the Indian Ocean, despite averaged time series are significantly correlated (Fig. 3) and the amplitudes of the leading EOF present the same variability (Fig. 10), the spatial patterns are clearly different. Therefore we have considered the two fields as non-consistent in this region.

4. Regional sea level changes during 1950–2003 and mass contribution

In those regions where steric-corrected altimetry and GRACE data are consistent at interannual time scales, the mass contribution to sea level changes during the second half of the 20th century may in principle be estimated by subtracting the steric contribution from total sea level. For past decades (1950–2003), total sea level is available through the reconstruction described in Section 2.1 (Llovel et al., 2009), which approaches altimetry from 1993 onwards. The reliability of the reconstructed fields is limited by the steadiness of the spatial patterns obtained during the altimetric period and by the uneven distribution of tide gauge stations. However, previous studies have demonstrated the ability of such methodology to capture the regional sea level variability both globally (Church et al., 2004; Llovel et al., 2009) and regionally (Calafat and Gomis, 2009; Calafat et al., 2010).

Steric sea level is obtained integrating the Ishii T and S climatologies down to 700 m depth for the period 1945–2006. The uncertainty in the steric component can be estimated from the uncertainties associated with the monthly T and S fields. In a first step, the error associated with the specific volume is computed by

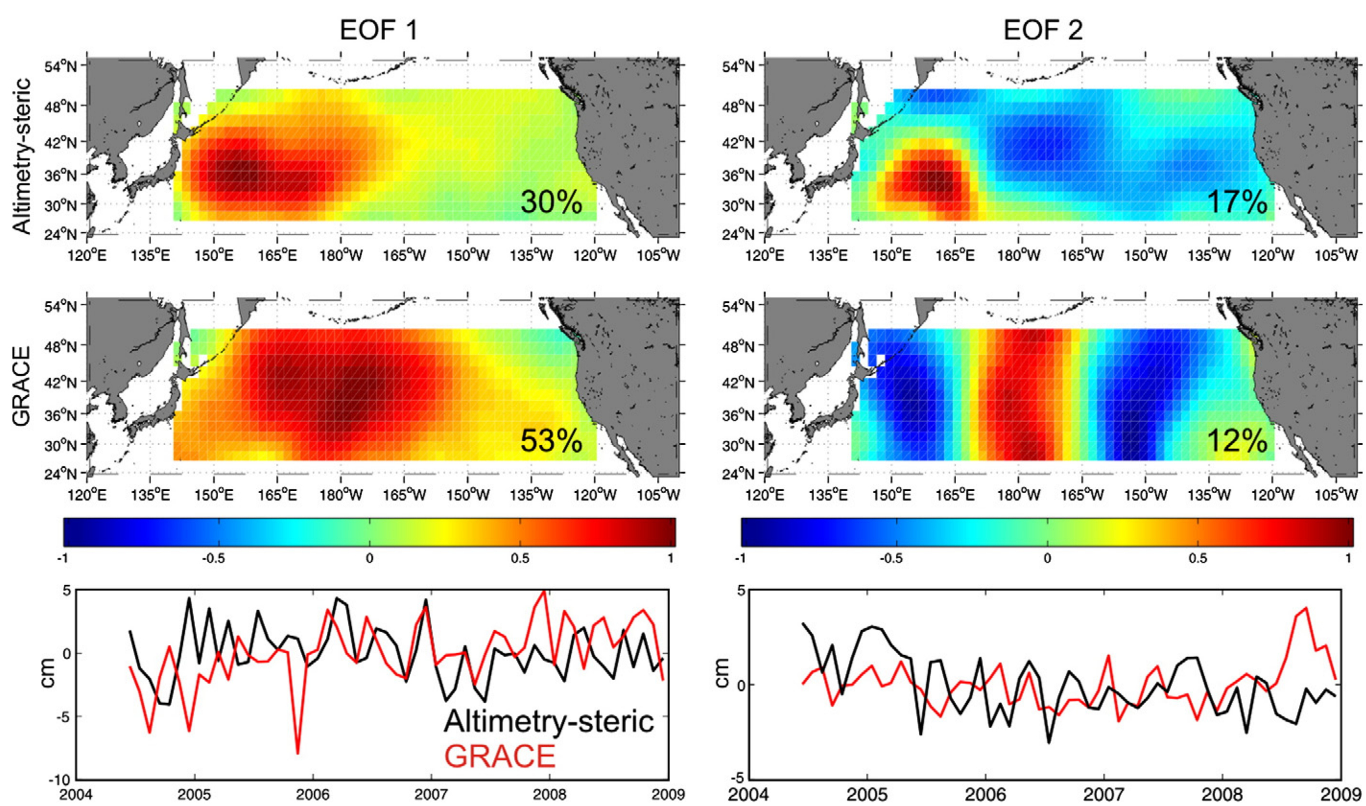

Fig. 6. First and second normalized EOFs of the northern Pacific decomposition for steric-corrected altimetry (top) and GRACE observations (middle). The corresponding amplitudes are shown in the bottom graphs.

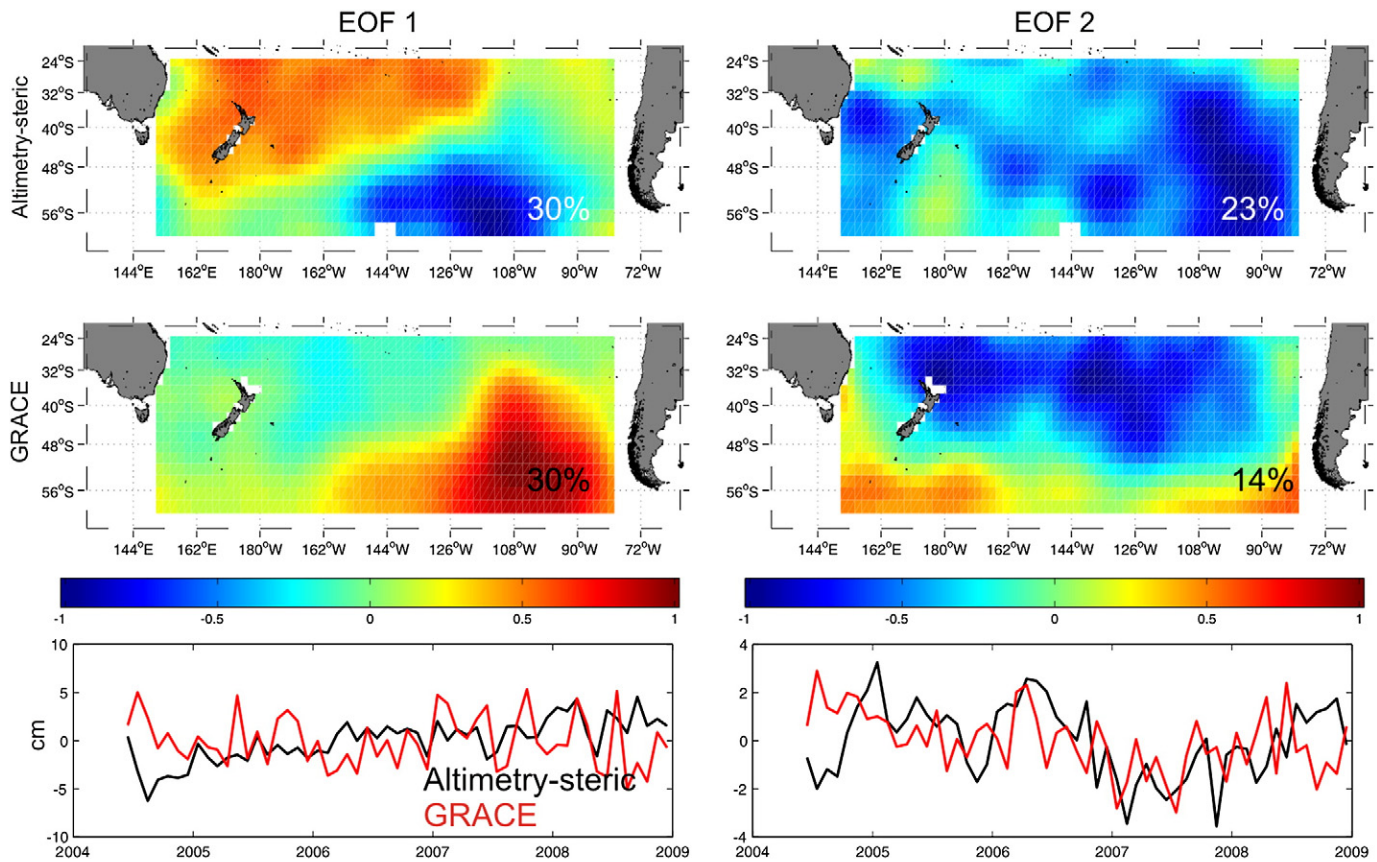

Fig. 7. As in Fig. 7, but for the southern Pacific region.

propagating the errors in T and S. To compute the error in the steric sea level we assume the worst scenario: that the error of the specific volume is vertically correlated and, therefore, the effect of the vertical integration is an error accumulation, rather than an error cancelation. The result will therefore be an upper boundary for the steric error. In a second step we estimate the error associated with the spatial mean steric sea level for each region, assuming in this case that errors are spatially uncorrelated; this is surely not true for small scales (adjacent grid points suffer from similar errors), but there is no reason to believe that errors are correlated at regional scale. More details on the methodology can be found in Calafat et al. (2010). Results yield typical error values between ± 0.7 and ± 1.6 cm for yearly regional averages, being larger at the beginning of the period, when observations are scarcer.

This methodology has of course some limitations. Firstly, those areas where it has been demonstrated that steric sea level estimated from interpolated data is not a good approximation must be discarded. This is the case of the North Atlantic and the Indian Ocean. Also the coverage and quality of measurements of the thermohaline properties of the ocean diminishes backwards in time. In particular, the interpolation of the scarce salinity measurements cannot be considered very reliable and thus only thermosteric changes can be accounted for. This in turn introduces further uncertainties in the steric estimation, but they are considered to have a small impact, since T changes dominate steric sea level variability everywhere except in the north Atlantic (Antonov et al., 2002). A further limitation comes from the fact that thermosteric sea level is integrated down to 700 m, which implies that the contribution of deeper layers to thermal expansion is neglected. Finally, the interpolation method for T data also introduces uncertainties, though they are provided for the Ishii climatology.

In order to account, as accurately as possible for these limitations, we have carried out a careful determination of linear trends and their associated uncertainties: linear trends are computed using an MM-regression estimator (Yohai, 1987), which is robust against outliers and allows including time-varying random errors. These random errors are in our case the uncertainties related to interpolation errors explained above. For more details see Appendix A.

If the errors associated with the variables have constant variance and there are no outliers in the data, then ordinary least squares (OLS) and robust estimators will lead to similar estimates for both the coefficients and the standard errors. However in the presence of errors having non-constant variance (heteroskedasticity), OLS will underestimate standard errors. Moreover if data also suffer from outliers, the coefficient estimates can be seriously biased. A robust standard error consistently estimates the true standard error even for data that suffer from heteroskedasticity and outliers. In order to illustrate this we have computed the thermosteric sea level trend for the North Pacific for the period 1945–2006 by means of both an OLS and an MM-regression estimator. For the OLS we have obtained a trend of -0.09 ± 0.04 mm/yr. In the case of the MM-estimator we have taken into account the uncertainties associated with the thermosteric sea level (which we know are larger at the beginning of the period, i.e., they suffer from heteroskedasticity). The thermosteric sea level trend obtained from the MM-estimator is -0.20 ± 0.05 mm/yr.

Regional sea level trends and their uncertainties for all regions except the north Atlantic and the Indian Ocean are listed in Table 2. Regional trends of total sea level vary between 1.5 and 1.7 mm/yr according to the sea level reconstruction. Values for thermosteric sea level are much smaller everywhere, ranging between 0.03 and 0.58 mm/yr. The remaining observed sea level rise is attributed to two factors: the thermal expansion of the deeper layers and the changes in ocean bottom pressure caused by mass variations. Regarding the contribution of the deep layers, Guinehut et al. (2006) used Argo data and sea level anomalies from altimetry to conclude that the differences between both data sets when steric sea level is computed with respect to a reference level at 700 m and at 1500 m is less than 10%.

It turns out, therefore, that the contribution of water mass changes dominates sea level changes in all regions. Our approach yields trends varying between 1.05 ± 0.07 mm/yr in the equatorial Atlantic and 1.57 ± 0.07 mm/yr in the north Pacific (Table 2). These values represent between 65% and 96% of the total observed regional sea level rise.

5. Discussion and final remarks

Comparisons among sea level from altimetry, steric sea level estimated from hydrographic data bases and ocean mass changes observed by GRACE have shown that the annual cycle of the ocean mass is in general well approximated by steric-corrected altimetry. Regionally averaged seasonal cycles are highly unsteady in time and represent only a small fraction of the total seasonality observed in sea level, in agreement with Llovel et al. (2010). At inter-annual scales the correlation between inferred and measured regional mass variations is smaller. We have also found that regional ocean mass variability is significantly larger than global changes, similarly to what happens with total and steric sea level. Besides the fact that steric-corrected altimetry has better resolution than GRACE observations, two other reasons have been identified for the weaker consistency between the two fields at inter-annual scales. The first one is related with the GRACE processing errors. The low signal-to-noise ratio of GRACE data prevents from making satisfactory comparisons with steric-corrected altimetry, in agreement with Llovel et al. (2010). This problem is at least partially overcome when working with regionally averaged sea level. We have found significant correlations in the north and south Pacific, in the south Atlantic and in the Indian Ocean. Conversely, results are not satisfactory in equatorial regions and in the north Atlantic. The second reason for the lack of consistency is the inability of interpolated T and S data to account for steric sea level in areas with large variability such as the Gulf Stream region.

The comparison of mass variations among different regions has revealed the exchange of ocean mass between northern and southern latitudes in the Atlantic and Pacific Oceans at inter-annual time scales. Furthermore such exchanges occur out of phase between the two oceans, although the short length of the GRACE time series prevents from drawing definitive conclusions with respect to the underlying mechanisms that drive this variability.

The consistency between regional steric-corrected altimetry and GRACE observations has been examined through the EOFs analysis and has revealed similar patterns of oscillation in the North and South Pacific and in the South Atlantic. In the latter moreover the barotropic mode of the Argentine basin is the main pattern in both data sets. The second EOF of GRACE data often reflects track-like patterns.

Linear trends of the mass contribution to sea level computed by GRACE data are highly dependent on the GIA correction applied. Further work is clearly needed to reconcile the currently available GIA

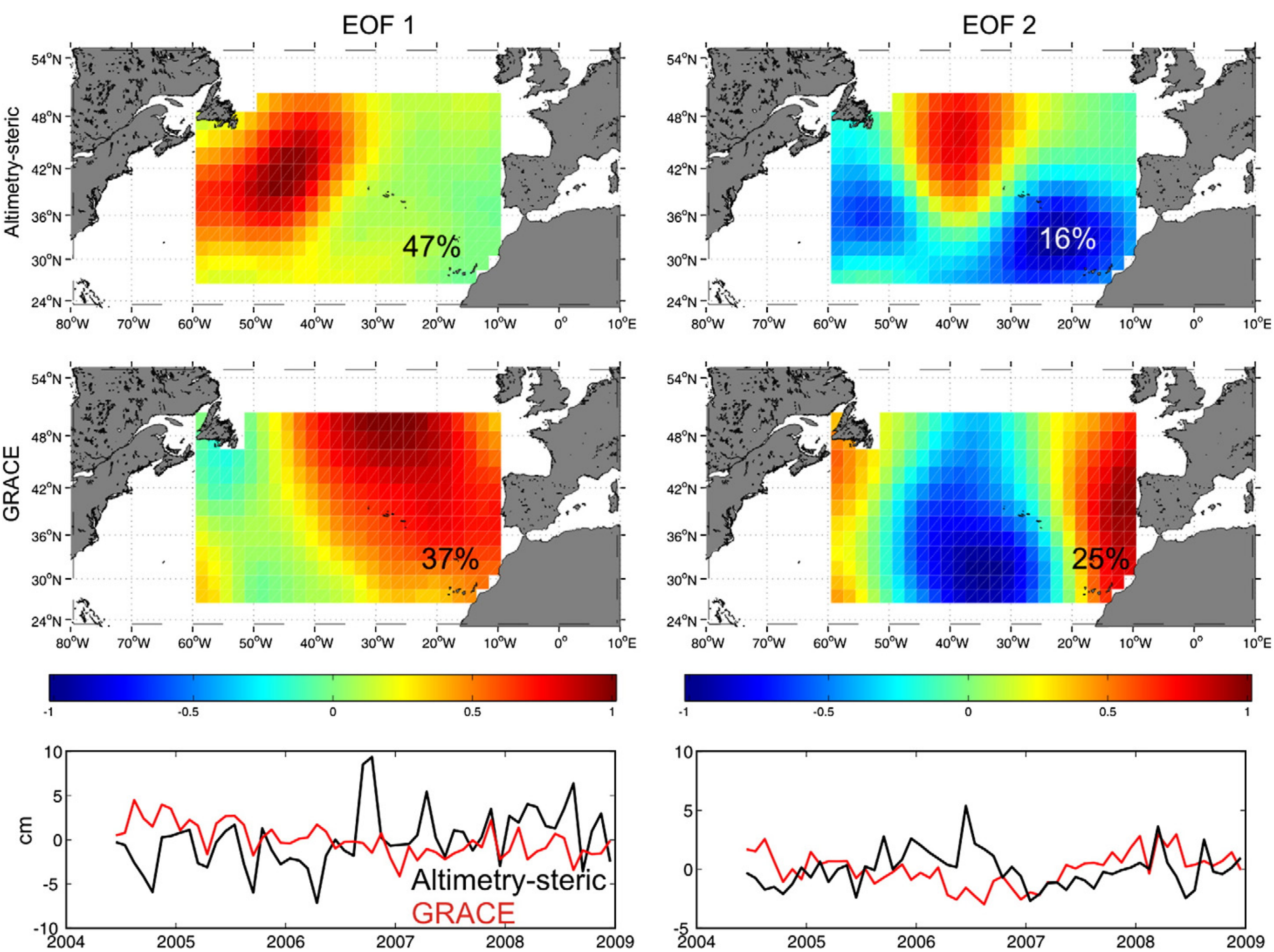

Fig. 8. As in Fig. 7, but for the North Atlantic region.

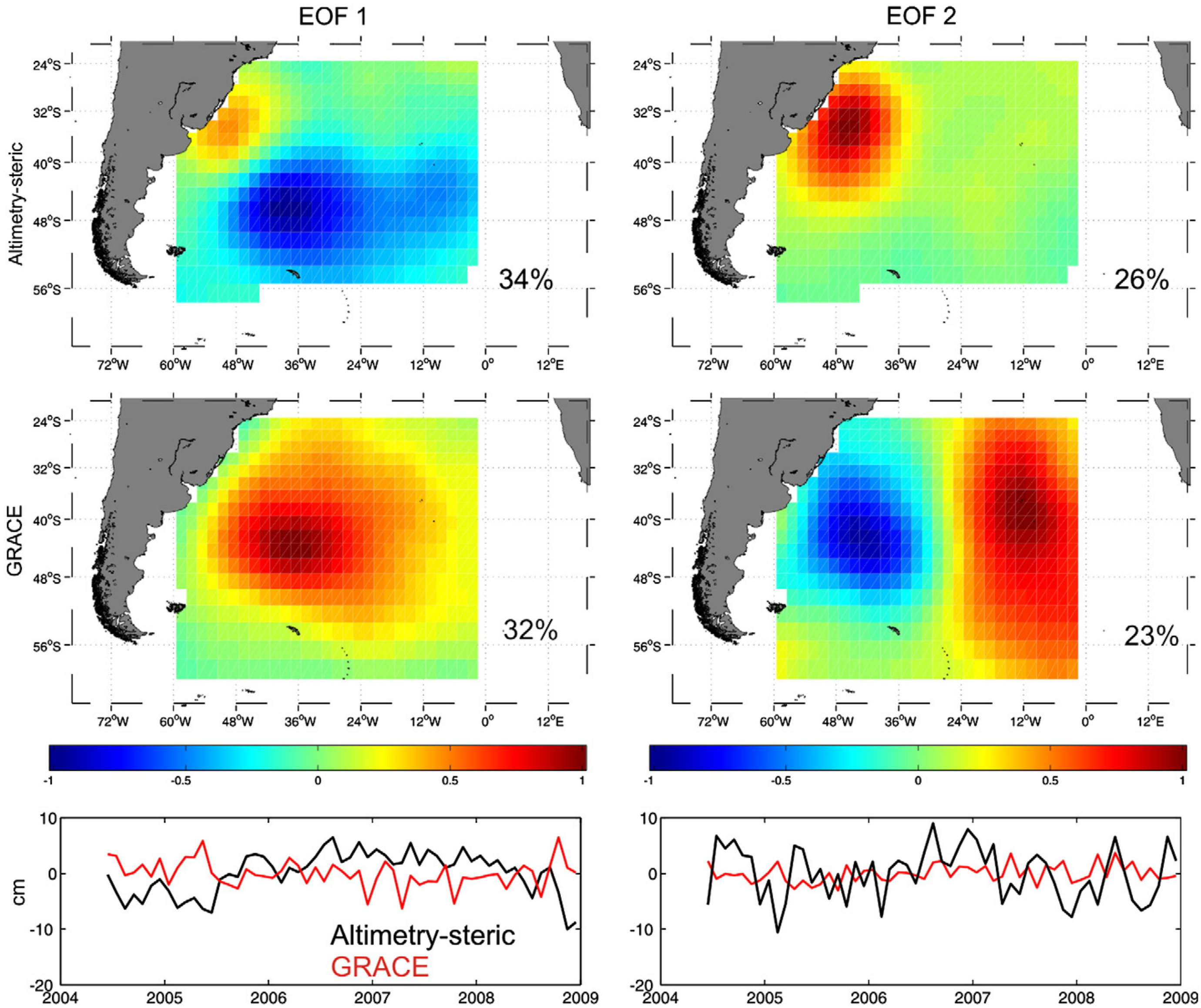

Fig. 9. As in Fig. 7, but for the South Atlantic region.

corrections provided by Paulson et al. (2007) and Peltier (2009). Regional sea level budgets cannot be closed using any of the two corrections, but the agreement is higher in most regions when using the correction chosen for this work (the one by Paulson et al., 2007).

Regarding longer term trends, comparisons of the thermal expansion of the top 700 m against total sea level rise for last decades indicates that the former is a minor contributor to the latter in all regions worldwide. Assuming that the thermal expansion of the layers deeper than 700 m is much smaller than that of the top layers, we conclude that the mass addition is the main contributor to regional mean sea level rise during the second half of the 20th century. This applies to all the regions examined, indicating that the origin of the observed mass increases is not a redistribution of ocean mass between regions, but a net global increase resulting from fresh water addition due to melting of glaciers and ice-sheets.

Our result is in agreement with Miller and Douglas (2004), who pointed at mass increase as the dominant factor in global mean sea level rise during the past century based on tide gauge observations and hydrography. Conversely, this result contrasts with the global average obtained by Domingues et al. (2008), who estimated a contribution of about 0.8 mm/yr of mass addition of a total sea level rise of 1.6 ± 0.2 mm/yr for the period 1961–2003. Their estimate of the thermosteric contribution of the upper 700 m is 0.52 ± 0.08 mm/yr, which is about 50% larger than the 0.31 ± 0.07 mm/yr given by Ishii et al. (2006) and the 0.33 mm/yr given by Antonov et al. (2005) for the period 1955–2003, also for the upper 700 m. Part of the disagreement may be caused by the fact that Domingues et al. (2008) assumed a linear increase in the rate of change of the contribution of Greenland and Antarctic ice sheets, despite the very little information available to constraint these values. Moreover, they used a deep-ocean thermosteric contribution of 0.2 mm/yr, that is, 40% of their estimate for the thermosteric contribution of the upper 700 m; this is in contradiction with the results obtained by Guinehut et al. (2006), who concluded that the contribution of the layers deeper than 700 m is much less important. If only the upper-ocean thermosteric contribution is taken into account, then the mass contribution is of the order of 1.1 mm/yr when using the estimate given by Domingues et al. (2008) and about 1.3 mm/yr when using the estimates given by Ishii et al. (2006) and Antonov et al. (2005). These estimates are in better agreement with our results.

When quantifying the mass contribution to long term regional sea level rise in terms of non-steric sea level, the computation of the regional steric component is a significant source of uncertainty. The other source is the reconstruction used to represent total sea level

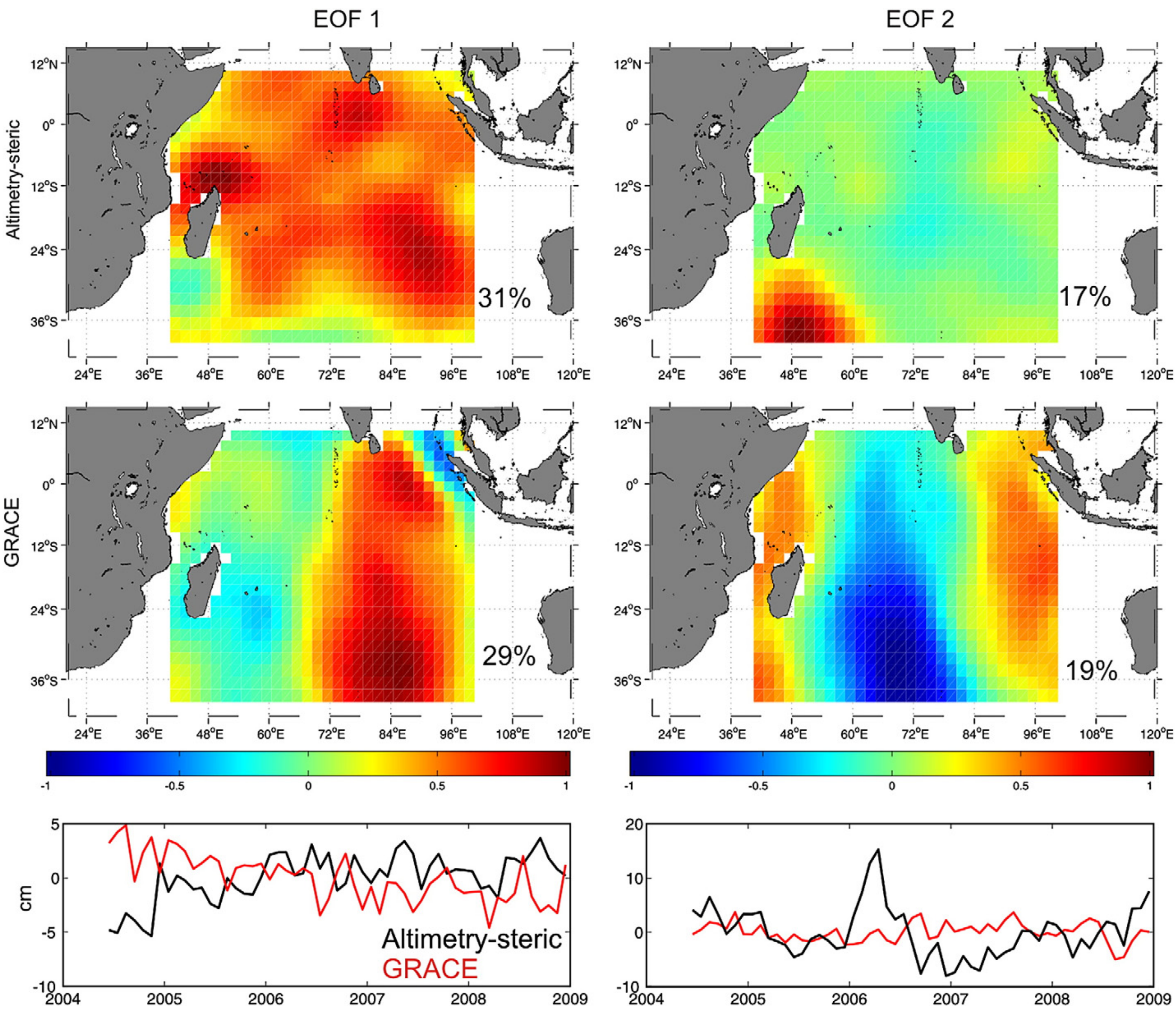

Fig. 10. As in Fig. 7 but for the Indian Ocean.

fields for the pre-altimetric period. Trends in reconstructed sea level are determined by the set of selected tide gauges and by the optimal interpolation method. Thus different spatial distribution of the tide gauge records can lead to small differences in regional sea level trends. Indeed, a region with many tide gauges will be strongly constrained by the optimal procedure to fit the tide gauge records, while regions with a sparse tide gauge distribution will be less constrained and can show differences, particularly at small scales. Further differences can be obtained depending on whether a full covariance matrix error is used or not for the interpolation. Despite all these uncertainty sources, however, the conclusion on the small fraction of the observed sea level rise accounted for by the thermosteric contribution remains unchanged.

An interesting question that remains open is whether the spatial pattern of the mass contribution to long term sea level rise can provide information on the sources of such fresh water input. Recently, Stammer (2008) derived the response of the ocean circulation to enhanced fresh water input associated with melting ice-sheets using an ocean general circulation model. He established that the dynamic response to ice melting implies the development of Kelvin and Rossby waves that propagate the sea surface height anomalies into the ocean basins at different time scales. According to Stammer (2008) results, the dynamic response would be much larger than the gravity response to the melting of continental glaciers and ice sheets. The latter induces spatial gradients of sea level due to the change of the geoid height, with lower than mean values close to the melting location and higher values in the far field (Mitrovica et al., 2001; Tamisiea et al., 2001). The linear trends of the mass contribution to sea level obtained for the period 1950–2003 are mapped in Fig. 11. Our results show striking similarities with the maps of sea surface height anomalies derived from the melting of Greenland (see Fig. 6 in

Table 2
Linear trends of total and thermosteric sea level and the difference between them.

	Reconstruction (total sea level)	Thermosteric	Total − thermosteric
N. Pacific	1.63 ± 0.04	0.03 ± 0.07	1.57 ± 0.07
Eq. Pacific	1.69 ± 0.04	0.24 ± 0.07	1.41 ± 0.08
S. Pacific	1.52 ± 0.03	0.14 ± 0.04	1.41 ± 0.04
Eq. Atlantic	1.63 ± 0.03	0.58 ± 0.07	1.05 ± 0.07
S. Atlantic	1.60 ± 0.02	0.22 ± 0.05	1.39 ± 0.05

Stammer, 2008). The coherence between both spatial patterns points at Greenland as the major source of fresh water input during the second half of the 20th century. However, given the limitations inherent to the interpolation of hydrographic data and to the use of reconstructed sea level field, further research is needed to ensure this point.

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Appendix A

Linear trends are computed using an MM-regression estimator (Yohai, 1987) calculated with an initial S-estimate (Roussseeuw and Yohai, 1984). The MM-regression estimator is computed with loss functions in Tukey's bi-square family. The tuning constants have been chosen to obtain simultaneous 50% breakdown-point and 95% efficiency when the errors are normally distributed. The S-estimate has been computed by means of the fast algorithm for S-regression estimates developed by Salibian-Barrera and Yohai (2006). While MM-estimators are robust against outlying observations and heteroskedasticity, standard errors estimates also need to be reliable, in the sense of not being overly biased by the presence of outliers and heteroskedasticity. In order to understand the importance of this point, let us consider the regression model

$$y_i = \vec{x}_i'\beta_0 + \sigma\varepsilon_i,\ i = 1,\ \ldots, n$$

where y_i are independent observations, $\vec{x}_i$ are the predictor variables, β_0 is the unknown regression coefficient to be estimated, ε_i are random errors, and n is the number of observations. Ideally, one would like to assume that the distribution of the data follows some specific symmetric distribution (F_0) such as the standard normal distribution. To allow for the occurrence of outliers and other departures from the classical model, we will assume that the actual distribution F takes the form

$$F = (1-\varepsilon)F_0 + \varepsilon\tilde{F}$$

where $0 \le \varepsilon < \frac{1}{2}$ and $\tilde{F}$ is an arbitrary and unspecified distribution. Under this assumption, the interpolation errors associated with the steric sea level (see Section 2.1) can be taken into account to obtain a robust estimate of the errors associated with the linear trends by simply adding an error term of the form $\tilde{\varepsilon}_i N(0,1)$, where $\tilde{\varepsilon}_i$ is the error associated with the ith observation and $N\,(0,1)$ is the standard normal distribution with mean 0 and unity variance. Steric sea level errors do not have constant variance, mainly due to the fact that the number of observations is larger at the end of the period than at the beginning, and therefore the actual distribution of y_i is of the form of F. In the cases that the errors $\tilde{\varepsilon}_i$ are not known we set them equal to 0, and therefore, the estimates of the standard errors are associated with natural variability and unknown random errors.

The standard error of robust estimates can be estimated using their asymptotic variances. However, the asymptotic distribution of MM-estimates has mainly been studied under the assumption that $F = F_0$, which does not strictly hold in many situations. In order to obtain robust estimates of the errors associated with the trends, we have used the fast bootstrap method proposed by Salibian-Barrera (2006), which yields a consistent estimate for the variance of the trend under general conditions. The simulation used to approximate the bootstrap distribution consists of bootstrapping the residuals of the MM-estimate.

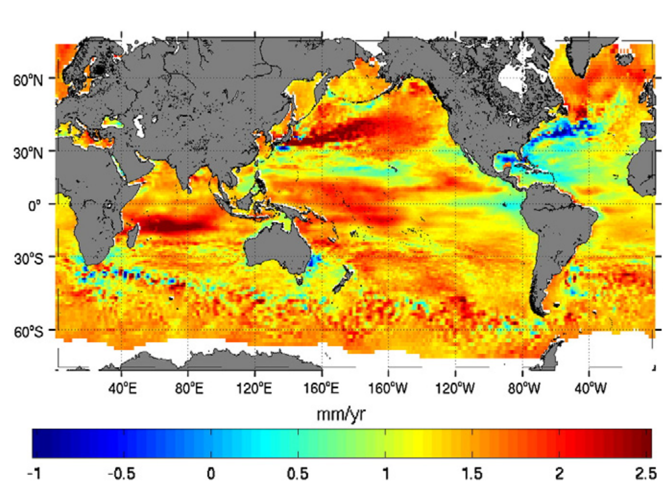

Fig. 11. Linear trends of the mass contribution to sea level rise for the period 1950–2003 inferred from the difference between reconstructed sea level and the thermosteric contribution.

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