Steric and mass-induced sea level variations in the Mediterranean Sea revisited

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[1] The total sea level variation (SLV) is the combination of steric and mass-induced SLV, whose exact shares are key to understanding the oceanic response to climate system changes. Total SLV can be observed by radar altimetry satellites such as TOPEX/POSEIDON and Jason 1/2. The steric SLV can be computed through temperature and salinity profiles from in situ measurements or from ocean general circulation models (OGCM), which can assimilate the said observations. The mass-induced SLV can be estimated from its time-variable gravity (TVG) signals. We revisit this problem in the Mediterranean Sea estimating the observed, steric, and mass-induced SLV, for the latter we analyze the latest TVG data set from the GRACE (Gravity Recovery and Climate Experiment) satellite mission launched in 2002, which is 3.5 times longer than in previous studies, with the application of a two-stage anisotropic filter to reduce the noise in high-degree and -order spherical harmonic coefficients. We confirm that the intra-annual total SLV are only produced by water mass changes, a fact explained in the literature as a result of the wind field around the Gibraltar Strait. The steric SLV estimated from the residual of "altimetry minus GRACE" agrees in phase with that estimated from OGCMs and in situ measurements, although showing a higher amplitude. The net water fluxes through both the straits of Gibraltar and Sicily have also been estimated accordingly.

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1. Introduction

[2] The mean sea level varies primarily in two ways: the steric sea level variations (SLV) produced by changes in the density and hence the volume of the water column, and the mass-induced SLV by addition/subtraction of water to/from the water column. The simple sum of these two constitutes the total SLV. The total SLV can be observed by satellite altimetry in the (absolute) terrestrial reference frame, marked in particular by the high-precision altimeter satellite TOPEX/POSEIDON (T/P) launched in 1992 and its follow-on missions Jason-1 and Jason-2. The steric SLV can be estimated from depth profiles of temperature and salinity from oceanic in situ measurements or from ocean general circulation models (OGCM), which often assimilate the said measurements. The mass-induced SLV have a gravitational signature and hence can be estimated from time-variable gravity (TVG) measurements, such as those available from the satellite mission of the Gravity Recovery and Climate Experiment (GRACE) launched in 2002.

[3] In order to understand sea level rise as a consequence of global climatic changes, it is desirable to identify the relative contributions of the steric and mass-induced terms in producing the observed SLV in space and time. The two contributions each accounted for about half of the global sea level rise of 3.1 mm/yr during 1993–2003 (http://www.ipcc.ch/ [*Cazenave and Nerem*, 2004; *Intergovernmental Panel on Climate Change*, 2007]), which represents an acceleration compared to the ~1.8 mm/yr for the entire 20th century estimated from tide gauge data [*Douglas*, 1997] or ~2.46 mm/yr if the artificial reservoir water impoundment is taken into consideration [*Chao et al.*, 2008]. However, during the TVG and Argo era [*Freeland et al.*, 2010] of 2003–2009, the long-term contribution of each component have been reported as 50% [*Leuliette and Miller*, 2009], around 80% for the mass-induced SLV [*Cazenave et al.*, 2008], and without closure between the two components and the total SLV [*Willis et al.*, 2008]. Further studies are necessary to conciliate those results.

[4] We shall combine data from satellite altimetry, GRACE TVG, in situ temperature and salinity profiles, and OGCM outputs in order to elucidate on the separation of steric and mass-induced SLV [*Chambers*, 2006a; *García et al.*, 2007a; *Lombard et al.*, 2007; *Willis et al.*, 2008; *Cazenave et al.*, 2008; *Leuliette and Miller*, 2009]. We apply this scheme to the Mediterranean Sea, a semienclosed basin where the mass-induced SLV are mainly produced via

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water exchange with the Atlantic Ocean through the Gibraltar Strait and with the atmosphere through precipitation and evaporation (in the form of $P - E$ or precipitation minus evaporation). Previous studies have already addressed this problem based on 2 years of GRACE data filtered using the classical isotropic Gaussian filter [*Fenoglio-Marc et al.*, 2006, 2007; *García et al.*, 2006, 2007b]. We revisit this problem using the latest releases of altimetry and GRACE data. The latter set with a 3.5 times longer time span, and applying nonisotropic filter which proved to be more appropriated for the GRACE data processing. Specifically, we shall examine the seasonal and nonseasonal mass-induced SLV from GRACE in relation to the total and steric SLV, the latter estimated from (1) OGCM outputs and in situ measurements and (2) the "altimetry minus GRACE" residual. We report, along with an apparent problem of amplitude estimation of steric SLV from different data sets, new findings w.r.t. Mediterranean Ocean circulation mass budget and its dynamical behavior.

2. Data Description and Processing

[5] Since its launch in March 2002, the GRACE mission has been measuring the gravity field to an accuracy that reveals the small TVG signal signifying mass transports on or in the Earth [*Tapley et al.*, 2004]. The data are routinely released as monthly sets of spherical harmonics, or Stokes coefficients [*Bettapdur*, 2007a]. We use the 78 monthly sets of the release 4 solution provided by the Center for Space Research (http://podaac-www.jpl.nasa.gov/grace/), spanning September 2002 to March 2009 (with missing June 2003, which we simply linearly interpolate). Some modeled geophysical signals were removed during the solution process: gravity perturbations due to the Sun, Moon and the other planets, solid Earth and oceanic tides (including pole tide), nontidal mass variability of the atmosphere and oceans, as well as secular trends of the C_{20}, C_{30}, C_{40}, C_{21}, and S_{21} coefficients [*Bettapdur*, 2007b].

[6] As we are interested in the Mediterranean Sea, the corrected atmospheric and ocean signals have been added back to the GRACE data (see *Chambers* [2006b] for more details). Then, the full ocean bottom pressure (OBP) variations are restored over the ocean [*Flechtner*, 2007]. The corrected secular variations have also been recovered. GRACE is not sensitive to the degree 1 (e.g., geocenter motion), but they should be added back to the GRACE; we use the time series estimated by *Swenson et al.* [2008]. Finally, we replace the C_{20} coefficient, which is poorly determined by GRACE, with the series derived from satellite laser ranging technique [*Cheng and Ries*, 2007].

[7] Assuming that the gravity variations are produced by mass variations on the surface of the Earth (such as the water mass transport within the water cycle, the biggest mass variations of the Earth in the intra-annual timescale), the surface mass variation in terms of the equivalent water thickness can be uniquely determined as [*Wahr et al.*, 1998; *Chao*, 2005]

$$\Delta\sigma(\theta,\lambda,t) = \frac{a\rho_E}{3}\sum_{n=0}^{\infty}\sum_{m=0}^{n}\frac{(2n+1)}{(1+k'_n)}P_{nm}(\cos\theta) \cdot [\Delta C_{nm}(t)\cos(m\lambda) + \Delta S_{nm}(t)\sin(m\lambda)], \quad (1)$$

where (θ, λ) = (colatitude, longitude), a and ρ_E are the equatorial radius and the mean density of the Earth, k'_n is the degree n load Love number, P_{nm} is the 4π-normalized associated Legendre function of degree n and order m, and $\Delta C_{nm}(t)$ and $\Delta S_{nm}(t)$ are the monthly Stokes coefficient anomalies from GRACE with respect to the 2003–2008 mean. Over the ocean, an increase of the surface mass of 1 kg/m^2 can be interpreted as an increase of OBP produced by an increase of 1/1.029 mm of sea level, where 1029 kg/m^3 is the mean density of seawater.

[8] The GRACE TVG data consist of Stokes coefficients up to degree and order 60. If they were noise free the expansion in equation (1) would be truncated at degree 60, hence a spatial resolution of about 20,000/60 ≈ 330 km. However, this is far from reality since it is known that GRACE solutions of high degree are plagued with noise [e.g., *Wahr et al.*, 2004], and the use of spatial filtering becomes crucial. Until recently, the most widely used filter, and the one used in previous GRACE-based studies of the Mediterranean, is the isotropic (Gaussian) filter [*Jekeli*, 1981; *Swenson and Wahr*, 2002]. More advanced filters have been devised since. In this study we revisit the Mediterranean Sea with improved GRACE data processing via the consecutive application of two independent filters. The first one removes some correlated errors in the coefficients of even/odd degree for a given order, which are produced by the polar orbit sampling of the satellites [*Swenson and Wahr*, 2006], with parameters fixed as by *Chambers* [2006b] to improve the estimated signal estimate over the oceans. Subsequently, we apply a second anisotropic filter based on the formal error filter by *Chen et al.* [2006], which is described as

$$W_{nm}^{C} = \frac{RMS\left(\Delta C_{nm}^{Synthetic}\right)^2}{RMS\left(\Delta C_{nm}^{Synthetic}\right)^2 + (SIG(\Delta C_{nm})\cdot k)^2}, \quad W_{nm}^{S} = \frac{RMS\left(\Delta S_{nm}^{Synthetic}\right)^2}{RMS\left(\Delta S_{nm}^{Synthetic}\right)^2 + (SIG(\Delta S_{nm})\cdot k)^2}, \quad (2)$$

where W_{nm}^C and W_{nm}^S are the weights to be used in the Stokes coefficients $\Delta C_{nm}(t)$ and $\Delta S_{nm}(t)$ in equation (1), $RMS(\Delta C_{nm}^{synthetic})$ and $RMS(\Delta S_{nm}^{synthetic})$ are the root mean square (RMS) of the Stokes coefficients of synthetic GRACE data set, $SIG(\Delta C_{nm})$ and $SIG(\Delta S_{nm})$ are the reported formal errors of GRACE data, and k is a parameter to be determined as explained below. The weights, W_{nm}^C and W_{nm}^S, are fixed to one when the result of equation (2) is larger than one. This filter depends on GRACE and synthetic GRACE data, and on the parameter k, but it is constant over time.

[9] We construct the synthetic GRACE data set composed of a hydrology model output over the continents and the OBP field from an OGCM over the oceans (see below for a description of both models). We do not use the formal errors of GRACE, but an estimation of them as explained by *Wahr et al.* [2006]. That is, each Stokes coefficient time series, $\Delta C_{nm}(t)$ and $\Delta S_{nm}(t)$, is least squares fitted by sinusoids with frequencies of 4, 2, 4/3, 1, and 1/2 years and 161/365.25 days. The latter frequency is to account for the aliasing produced by the undersampling in GRACE monthly data of the S_2 oceanic and atmospheric tides. Then, the fitted signals are subtracted from the original data and the residual is con-

sidered as an overestimation of GRACE errors. The parameter k is estimated to maximize the ratio of signal variance between the continents and the oceans. When estimating the signal over the ocean, we do not consider the points closer than 750 km to the coasts, in order to avoid land contamination. In this case a value of $k = 0.7$ is obtained, in contrast to *Chen et al.* [2006] who found a value of k larger than 1 based on an underestimation of the errors.

[10] Figure 1 demonstrates the strength of the two-stage anisotropic filter. Figure 1a shows a snapshot of the synthetic GRACE data in the Mediterranean Sea and surrounding areas for October 2004 and May 2005. These grids are developed in a Stokes expansion, and then reconstructed using coefficients up to degree 50 [*Heiskanen and Moritz*, 1967]. The results are shown in Figure 1b; due to the truncation these maps show lower amplitudes as well as smaller spatial resolution. However, this is the best possible reconstruction of the signal using Stokes coefficients up to degree 50 to be used as a reference of quality. In order to see the effect of our anisotropic filter in the signal, we apply it to the data before the reconstruction. As expected, Figure 1c shows lower amplitude and spatial resolution than the truncated maps. In comparison, the same procedure applied using the isotropic filter with a radius of 750 km, which would remove almost all the north-south stripes in GRACE data, leads to even lower amplitude and smaller spatial resolution (Figure 1d). Furthermore the leakage from the continental signal becomes more evident, even changing the sign of the signal in some regions of the Mediterranean with respect to the truncated nonfiltered signal.

[11] From this experiment it is clear that it is impossible to avoid leakage of continental hydrology when using a truncation at degree 50, regardless of any filter, except if the continental hydrology is forward modeled prior to the inversion of GRACE data [*Sabaka et al.*, 2010]. In particular, the signal in the Adriatic and Aegean seas is significantly contaminated by continental signals, so we exclude them from the computations over the Mediterranean Sea. Figure 2a shows the GRACE signal averaged over the Mediterranean (excluding the cited seas), from real GRACE data filtered with our anisotropic filter (red line), compared favorably with the isotropic filter (black line). This signal includes continental leakages, which is reduced as follows. The Mediterranean signal is set to zero in the anisotropic-filtered GRACE maps, expanded into Stokes components and then reconstructed only up to degree 50. A spurious signal appears in the Mediterranean, as an estimate of leakage of the continental signal. This estimate is subtracted from the GRACE maps over the Mediterranean to reduce the continental leakage [*Wahr et al.*, 1998; *Chambers*, 2006b]. Figure 2b shows the average leakage signal; subtracting it from the averaged GRACE signal in Figure 2a produces the residual in Figure 2c, which represents the Mediterranean OBP variation "without" continental leakage.

[12] On the other hand, from the experiments with the synthetic GRACE data it is observed that both truncation and filtering produce reduction of the signal amplitude, which needs to be restored. A usual method consists in applying the selected filter to a grid with unity in the region of interest and zero elsewhere, then, the grids are reconstructed truncating the Stokes expansion at the same degree than the GRACE data. The mean value of the resultant grids over the region is assumed as the standard deviation amplitude reduction factor by the filter [*Velicogna and Wahr*, 2006; *Swenson and Wahr*, 2007]. This value is 0.46 for the Mediterranean Sea. However, the power spectral distribution of the unity grid is far from the one of GRACE data. In estimating a more realistic amplitude reduction factor, our anisotropic filter is applied to the synthetic GRACE data set described above, from which we found 0.56. Thus, the GRACE-derived signal is multiplied by 1/0.56 to restore the lost amplitude in all the following results.

[13] In order to transform the GRACE OBP into mass-induced SLV, the atmospheric pressure averaged over the global ocean must be subtracted [*Willis et al.*, 2008]. The OBP of a column of water tends to keep constant, so when the atmospheric pressure increases (decreases) the OBP, an amount of water accounting for the same pressure is evacuated from (added to) the column of water. This is the inverted barometer (IB) response of the ocean. So, OBP variations can be interpreted as mass-induced SLV not related to the IB response of the ocean. However, the latter is only true when there is no physical impediment to the IB response, as there is, for instance, when atmospheric pressures variations occur over the whole ocean basin, or over any closed basins. And that is the reason for the atmospheric pressure subtraction above. Although the Mediterranean Sea is a semienclosed basin, the response of the sea level to atmospheric pressure changes differs from a pure IB response at frequencies higher than (30 days)$^{-1}$ [*Le Traon and Gauzelin*, 1997]. So, as GRACE-derived, mass-induced SLV time series are monthly, they do not show the IB response of the ocean. We use the atmospheric pressure data from the JRA-25 Reanalysis [*Onogi et al.*, 2007], with spatial resolutions of 2.5° × 2.5° at monthly intervals.

[14] Besides the GRACE TVG data, we use the altimetry measurements from a merged solution of multiple radar altimetry satellites for the total SLV from January 2002 to November 2008. This solution combines T/P and ERS-1 or ERS-2 for the period January 2002 to August 2002, Jason-1 and ERS-2 for August 2002 to June 2003, Jason-1 and Envisat for June 2003 to January 2009, and Jason-2 and Envisat for January 2009 to March 2009. These data are the reprocessed delayed time (v3.0.0) mean sea level anomaly, which are given on grids at 7 day sampling intervals, with all the usual corrections applied, including the inverted barometer response of the ocean. This data set includes several improvements with respect to previous delayed time solutions. For example, the processing of the altimetric measurements near the coast has been enhanced, which is of great importance for the study of semienclosed basins such as the Mediterranean Sea (see www.aviso.oceanobs.com for more information). To make these data comparable to GRACE data, the altimetric data have been monthly averaged, and the grids have been reduced to 1° regular grids.

[15] Model outputs are also used in our study. They include oceanic temperature and salinity profiles, and OBP from two OGCM:

[16] 1. ECCO (Estimating the Circulation and Climate of the Ocean, http://www.ecco-group.org/products.htm [*Stammer et al.*, 2002]) model version kf080 is run on 1° (longitude) and 1–0.3° (latitude) grid, and forced by NCEP (National Centers for Environmental Prediction) Reanalysis fields. Sea

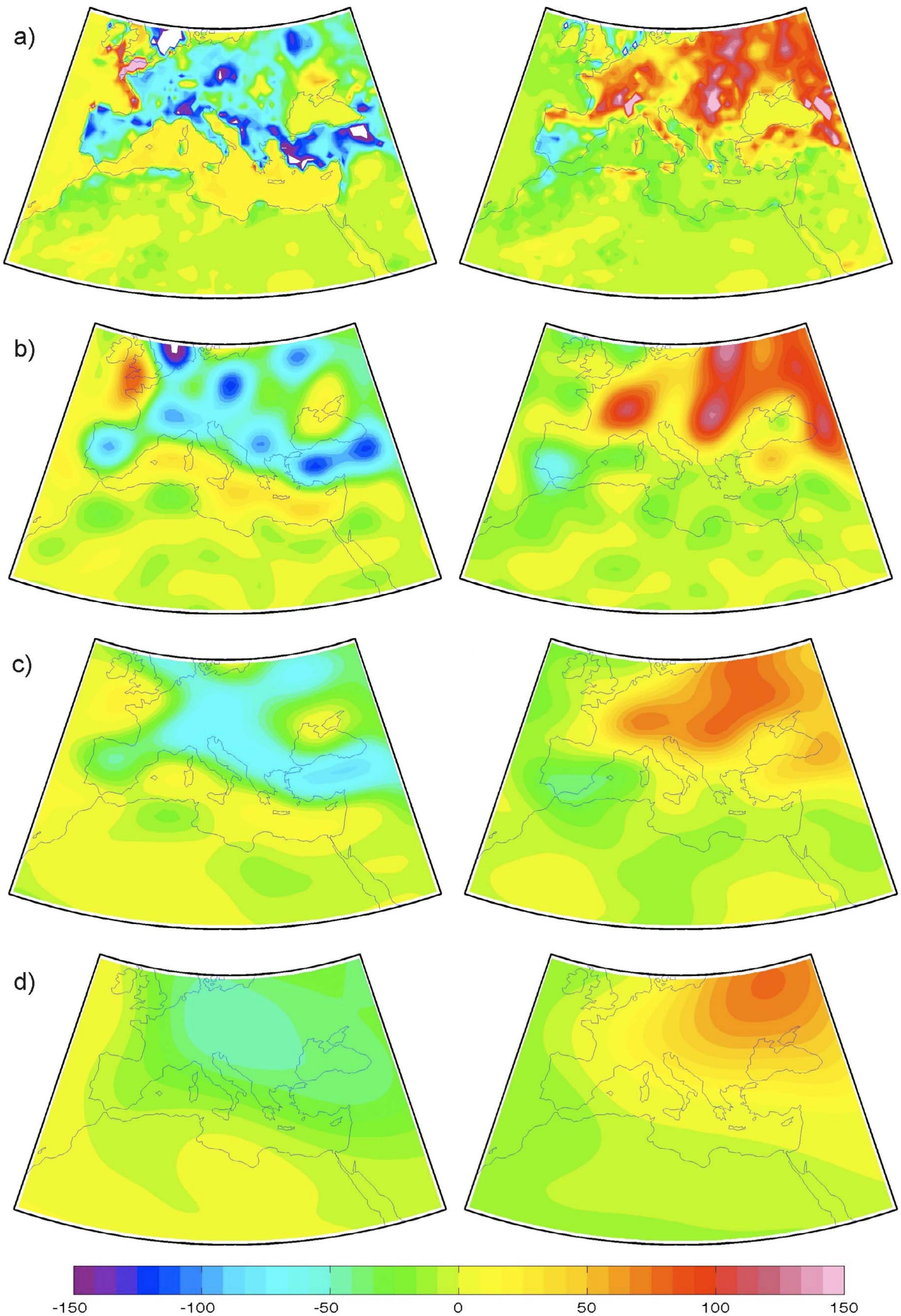

Figure 1

surface topography from altimetry and temperature, as well as XBT and Argo temperature profiles are assimilated via a Kalman filter. OBP output from ECCO are used to form the synthetic GRACE data.

[17] 2. MERCATOR model (the PSY3V1 version, see http://www.mercator-ocean.fr/ for more details [*Drévillon et al.*, 2008]) forced by the ECMWF (European Centre for Medium-Range Weather Forecasts) operational fields, and only assimilating altimetry data. The MERCATOR outputs are given at 0.5° regular grids at daily intervals, but only spanning the period of October 2005 to September 2008. All products have been averaged on 1° × 1° regular grids and to monthly samples to match those of the GRACE data. Both models are forced by atmospheric models, including net atmospheric freshwater fluxes, but not water mass exchange with the continents (e.g., river runoff), and they conserve volume instead of mass. In addition, we use temperature and salinity profiles from the EN3 v2a (http://hadobs.metoffice.com/en3/ [*Guinehut et al.*, 2009]) data set, which are monthly 1° × 1° regular grids optimally interpolated from in situ measurements belonging to WOD05, GTSPP, Argo, and ASBO data sets. Besides, we use for reference the historical climatology data from the World Ocean Atlas 2005 (WOA05, http://www.nodc.noaa.gov/OC5/WOA05/pr_woa05.html)

[18] The terrestrial water storage fields we use are derived from the GLDAS/Noah (Global Land Data Assimilation Systems) model (http://disc.sci.gsfc.nasa.gov/hydrology/data-holdings [*Rodell et al.*, 2004]). The 3 hour, 0.25° (no data southward of 60°S), and four-layer soil moisture, snow equivalent height and canopy water have been averaged to the same GRACE format as above for the period January 2002 to December 2008. These data are used to construct the synthetic GRACE data over the continents.

[19] We also use the data fields of the net vertical water flux in the atmosphere in the form of $P - E$, the precipitation minus evaporation, for the Mediterranean Sea. $P - E$ can be derived from the water vapor fluxes (Q) and the total water content of the atmosphere W [*Oki et al.*, 1995]

$$P - E = -\nabla_H Q - \frac{\partial W}{\partial t}. \quad (3)$$

Q and W are provided from the JRA-25 Reanalysis [*Onogi et al.*, 2007], with spatial resolutions of 1.125° at 6 hour intervals. We average $P - E$ estimates to monthly samples, for the period of January 2002 to March 2009.

[20] All these data sets are quite different in terms of their spatial and temporal sampling. Although they have been reduced to 1° monthly grids, they may not represent the same physical processes. In order to study the comparability between them, the spatial autocorrelation function of each data set is studied in the Mediterranean Sea, excluding the Adriatic and the Aegean seas, following *Dobslaw and Thomas* [2007] (see Figure 3a). Data with lower autocorrelation values are interpreted as data with better spatial resolution. ECCO and Mercator show similar autocorrelation functions for OBP and for steric SLV. The former should be similar to the autocorrelation function from GRACE OBP, and the latter to the one from steric SLV from EN3. However, the models show a more uniform signal. There are two possible explanations for these discrepancies: (1) As GRACE and EN3 are observations, they may measure some physical processes not modeled in the models. (2) The lower autocorrelation in the observations may also be produced by noise. In the case of GRACE, the noise may come from the imperfection of the filter and from the Gibbs effect produced in the continental leakage reduction process; and in the EN3 data sets it may be produced by the nonuniform distribution of the temperature and salinity profiles used to optimally interpolate the 1° grids. On the other hand, the autocorrelation function from altimetry is similar to those from the models though with lower values. So, the conclusion from this analysis is that the different data sets are not spatially comparable. However, they are comparable as far as the averaged time series over the basin are concerned. The reason is that averaging any grid in any basin implies a filtering process in the spectral domain, since each degree n order m Stokes coefficient of the spectral domain of the grid is weighted by the degree n order m Stoke coefficient defining the basin during the averaging process [*Swenson and Wahr*, 2002]. In order to illustrate the effect of the averaging process in the spectral power of the grids, we filter all data sets with a "Mediterranean filter" as follows: (1) A 1° regular grid with value one in the Mediterranean Sea and zero otherwise is expanded into Stokes coefficients [*Heiskanen and Moritz*, 1967]. (2) The latter are normalized to values between −1 and 1, and then the "Mediterranean filter" arises. (3) All data sets are expanded into Stokes coefficients, which are then weighted with the "Mediterranean filter" (simulating the averaging process in the spectral domain). (4) The grids are reconstructed from the filtered Stokes coefficients. Figure 3b show the spatial autocorrelation function of those grids, which are perfectly comparable (note the y axis values), as far as the averaged signal over the Mediterranean is concerned.

3. Results

3.1. Seasonal SLV

[21] The mean mass-induced SLV and OBP from GRACE data (filtered, leakage reduced, amplitude restored) are given in Figure 4 (red line) and studied below. They are nearly identical to the time series associated with the first Empirical Orthogonal Function (EOF, not shown), which accounts for 73% of the total SLV variance [*Preisendorfer*, 1988] and has a spatial pattern of positive everywhere, meaning a variation in phase of the whole Mediterranean basin.

Figure 1. (a) Raw synthetic GRACE data (GLDAS and ECCO OBP); (b) synthetic GRACE data are expanded into spherical harmonics and then reconstructed truncating the expansion at degree 50; (c) reconstruction of synthetic GRACE data after filtering with the two-stage anisotropic filter; (d) as in Figure 1c but synthetic GRACE data are filtered with a Gaussian filter of radius 750 km for (left) October 2004 and (right) May 2005. Units are mm of water thick equivalent.

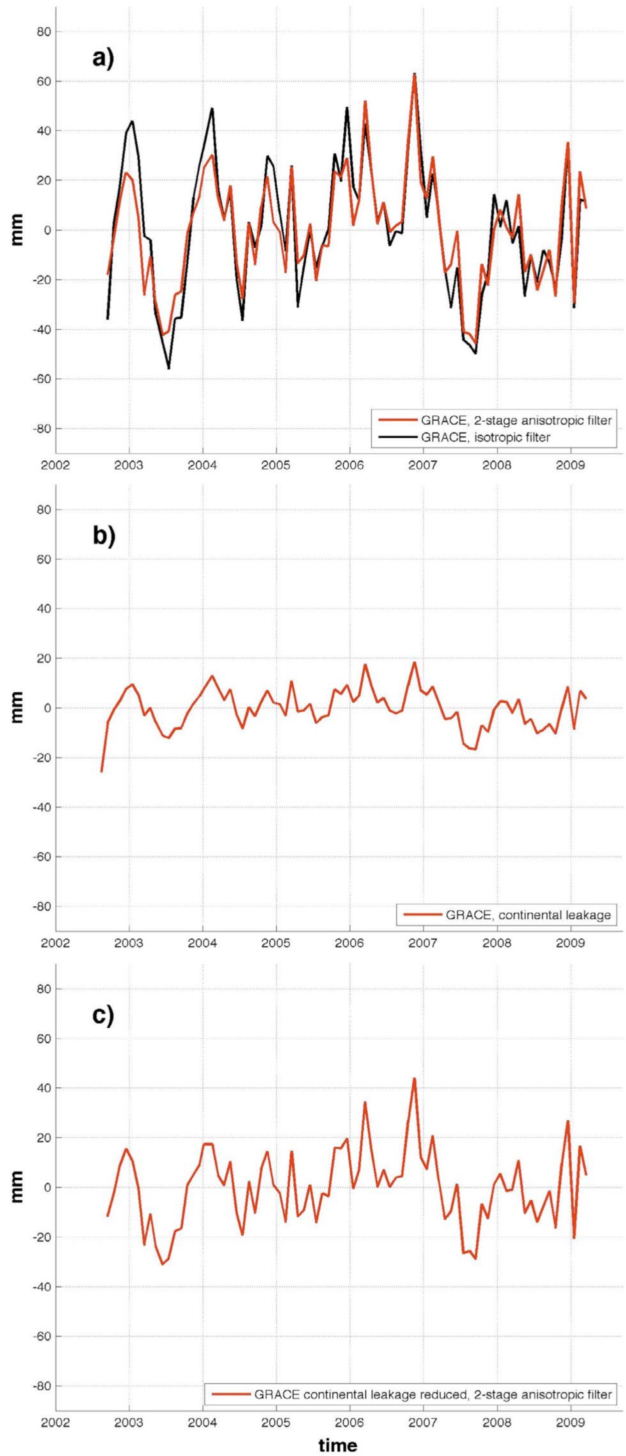

Figure 2. (a) GRACE signal averaged over the Mediterranean Sea after filtering by the two-stage anisotropic and the isotropic (Gaussian, r = 750 km) filters. (b) Leakage signal in the Mediterranean from the continents estimated from two-stage anisotropic-filtered GRACE. (c) Residual from the previous time series (Figures 2a and 2b). No amplitude restoring factor has been applied.

[22] To examine the seasonal and nonseasonal (intra-annual) signals in time series, we adjust by least squares fitting the time series to the following expression:

$$Signal \approx p_0 + p_1 \cdot t + p_2 \cdot t^2 + A_a \cos(\omega_a \cdot t - \phi_a) + A_{sa} \cos(\omega_{sa} \cdot t - \phi_{sa}), \quad (4)$$

where t represents time, (ω, A, ϕ) = (frequency, amplitude, phase) and the suffix a and sa denotes annual and semiannual terms, respectively. The obtained seasonal signals for the GRACE time series above have an annual amplitude and phase of 21 ± 7 mm and 4° ± 19° (peak on 4 January) for the OBP, and 29 ± 7 mm and 4° ± 13° for the mass-induced SLV, where the errors account for 95% of the confidence of the estimates (Table 1). The semiannual signal is not significantly different from zero.

[23] The observed OBP from GRACE is compared with the OBP from ECCO and Mercator OGCM, represented as mm of equivalent water thickness (Figure 4). Both data sets have been spatially averaged in the same way as GRACE, excluding the Adriatic and Aegean seas. The resultant curves are fitted to equation (4) and the obtained annual amplitudes and phases are: 38 ± 6 mm and 15° ± 9° (peak on 15 January) for ECCO OBP, 30 ± 13 mm and 14° ± 24° (peak on 14 January) for Mercator OBP (errors are larger because of the shorter time span), see Table 1 and Figure 5. Both OBP estimates agree well with GRACE in phases, but are larger than GRACE in the amplitude. The latter may or may not be indicative judging from the large seasonal variability during the period in Figure 4. Similar to GRACE data, model signals do not show a semiannual signal significantly different from zero.

[24] The Mediterranean mean SLV, as observed from altimetry and computed over the same grid points as the GRACE data, is given in Figure 6, along with their seasonal signals (after fitting to equation (4)). The dominant seasonal signal in altimetry is clearly annual with an amplitude of 72 ± 6 mm (about 2.5 times bigger than GRACE) and a phase of 273° ± 5° (peak around 4 October, about 3 months ahead of GRACE). Their large differences (see Figure 5) should signify a large steric contribution to SLV.

[25] The steric SLV are produced by density changes in the column of water, and we will estimate them in 2 ways: (1) from salinity and temperature profiles as derived, for example, from ECCO and Mercator models, from optimally interpolated in situ measurements from EN3 data set, or from the WOA05 climatology (for only the seasonality) and (2) as the residual by the simple subtraction of the mass-induced SLV from the altimetry-observed SLV [*Chambers*, 2006a; *Lombard et al.*, 2007; *García et al.*, 2007a] (hereafter referred as "altimetry minus GRACE"). Figure 7 shows the steric SLV estimates from all these possibilities. The steric annual amplitude and phases are (78 ± 5 mm, 250° ± 4°) for "altimetry minus GRACE," (42 ± 2 mm, 258° ± 2°) for ECCO, (68 ± 5 mm, 255° ± 5°) for Mercator, (58 ± 5 mm, 257° ± 5°) for EN3, and (67 ± 6 mm, 246° ± 5°) for WOA05 (see Table 1). The agreement in phase is remarkably good (peak between 6 and 19 September). Mercator amplitude is in good agreement with historic estimates from WOA05, and both of them agree with "altimetry minus GRACE" within the error estimates. However, EN3 and ECCO are

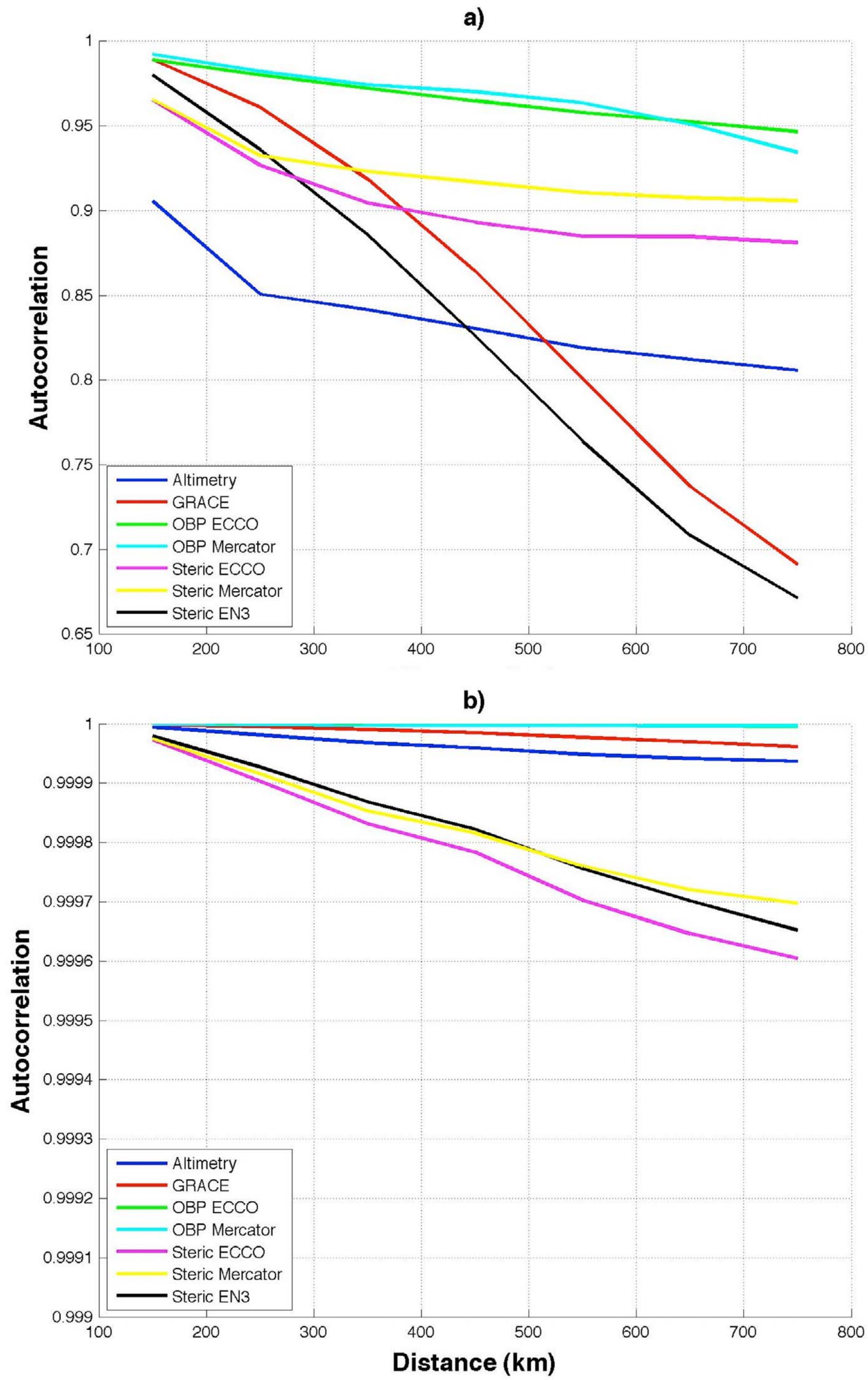

Figure 3. (a) Spatial autocorrelation functions over the Mediterranean Sea of different data sets. "GRACE" refers to mass-induced SLV. (b) Same as Figure 3a but after the "Mediterranean filter" has been applied.

considerably smaller than that of "altimetry minus GRACE," reaching only the 54% and the 74% of the latter (see Figure 5). The factor of underestimation, with respect to "altimeter minus GRACE," can be readily obtained in a least squares sense (using the total signal, not just the seasonal signal): 1.65 for ECCO, and 1.12 for EN3. These discrepancies in the amplitude can produces important differences when estimating the mass-induced SLV by forming the residual of "altimetry minus steric." As seen from Table 1, the amplitudes of those residuals agree within the error estimates with GRACE result, but their phases range are 291° for ECCO (peak on 22 October), and 318° for EN3 (peak on 18 November). Note that in the case of Mercator and WOA05,

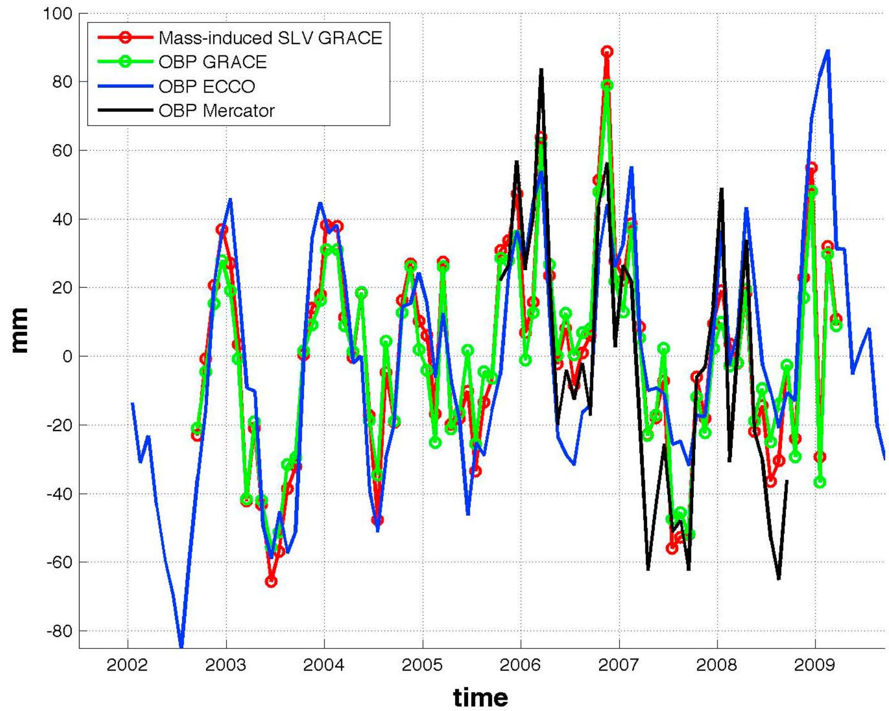

Figure 4. Mean time series for the mass-induced SLV and OBP of Mediterranean Sea, excluding Adriatic and Aegean seas. GRACE-derived times series are filtered by the two-stage anisotropic filter, continental leakage reduced, and amplitude multiplied by the restoring factor 1/0.56. The signals are mass-induced SLV from GRACE (red line), OBP from GRACE (green line), OBP from ECCO (blue line), and OBP from Mercator (black line).

the residual phases agree within the one from GRACE within the error estimate.

3.2. Nonseasonal SLV

[26] The seasonal and trend signal (represented by the right side of equation (4)) in the mass-induced SLV from GRACE has a standard deviation of 22 mm, while the nonseasonal residual, after subtracting the former from the original, has a standard deviation of 21 mm (see Figure 6). The corresponding nonseasonal SLV from altimetry has a comparable standard deviation of 20 mm. In fact, Figure 6b shows that the two signals are quite similar; their (broadband) cross-correlation coefficient is 0.71. This implies little variance is associated with the nonseasonal steric SLV.

[27] The intraseasonal signals observed in GRACE (both OBP and mass-induced SLV) are also observed by the models, despite smaller amplitudes (especially in 2005; see Figure 4). Both GRACE and OBP from ECCO show a positive trend up to 2006 and negative subsequently. Mercator also reproduces the latter behavior after 2006. This behavior is not observed in total SLV from altimetry (see Figure 6), since it is balanced by the interannual steric SLV, which show negative trend up to 2006 and positive subsequently (see Figure 7).

3.3. Water Mass Flux

[28] The total water mass flux (F) in the Mediterranean Sea is produced via the vertical flux of $P - E$, and the horizontal fluxes of river discharge (R), exchange with the Black Sea through the Bosphorus and Dardanelles straits (B) and with the Atlantic Ocean through the Gibraltar Strait (G)

$$F = P - E + G + B + R, \qquad (5)$$

where positive (negative) values represent gain (loss) of water by the Mediterranean. For a Mediterranean area of 2.5×10^{12} m^2 (including the Aegean and Adriatic seas), a mean sea level rise (drop) of 1 mm/month is equivalent to a net inflow (outflow) of 0.09 Sv/100 (1 Sv = 10^6 m^3 s^{-1}; and 1 Sv/100 = 10^4 m^3 s^{-1}). The variation of the Mediterranean water mass budget has already been estimated from GRACE data (Figure 4); F is simply its month-to-month derivative with units of mm/month (Figure 8). The total mass flux shows an annual signal with an amplitude of 1.4 ± 0.8 Sv/100 and a phase of 290° ± 34° (peak around 21 October).

[29] The $P - E$ signal averaged over the Mediterranean Sea can be seen in Figure 8. It is always of negative value (that is, always losing mass through vertical fluxes) with a mean value of −6.3 Sv/100, and estimated annual signal of (2.7 ± 0.4 Sv/100, 56° ± 8°), which is in agreement with historical estimates [*Boukthir and Barnier*, 2000; *Mariotti et al.*, 2002]. This flux of mass is mainly balanced (around 85%, see below) by water coming into the Mediterranean from the Atlantic Ocean through the Gibraltar Strait. This balance is not achieved instantaneously, giving rise to the annual variations in the Mediterranean mass budget (Figure 4).

Table 1. Annual Amplitudes and Phases (Indicating the Peak Time in the Year) of Some Mean Signals in the Mediterranean Sea

	Signals	Annual Amplitude (mm)	Annual Phase (°)	Annual Peak
Total SLV	Altimetry	72 ± 6	273 ± 5	4 October
Steric SLV	Altimetry (GRACE)	78 ± 5	250 ± 4	10 September
	ECCO	42 ± 2	258 ± 2	19 September
	Mercator	68 ± 5	255 ± 5	16 September
	EN3	57 ± 5	256 ± 5	17 September
	WOA05	67 ± 6	246 ± 5	6 September
Mass-induced SLV	GRACE	29 ± 7	4 ± 13	4 January
	Altimetry (1.65*Steric ECCO)	18 ± 7	345 ± 23	16 December
	Altimetry (Steric ECCO)	33 ± 6	291 ± 11	22 October
	Altimetry (Steric Mercator)	13 ± 13	1 ± 57	1 January
	Altimetry (Steric EN3)	24 ± 9	315 ± 21	18 November
	Altimetry (Steric WOA05)	32 ± 7	339 ± 13	10 December
OBP	GRACE	21 ± 7	4 ± 19	4 January
	ECCO	38 ± 6	15 + 9	15 January
	Mercator	30 ± 13	14 ± 24	14 January

[30] Given the above, the Gibraltar flux G can be estimated using equation (5), neglecting B and using a climatological cycle for R, the latter having a small annual signal and a mean value of 0.95 Sv/100 estimated by *Boukthir and Barnier* [2000] from 20 years of historical records (Figure 8). The resultant estimate of G is given in Figure 8, which shows positive values around 10–11 months during a year with a mean value of 5.5 Sv/100, and an annual signal (3.9 ± 0.8 Sv/100, 253° ± 14°) (peak around 13 September). See Table 2 for a summary of the annual signals of most of the components in equation (5).

[31] The same scheme can be applied to, say, the subbasin of the eastern Mediterranean Sea. Thus, knowing the $P - E$ and GRACE TVG of the region, one can estimate the in-out mass flux through the Strait of Sicily. We obtain for the latter an annual signal of (2.3 ± 0.5 Sv/100, 267 ± 12°) and a mean value of 4.4 Sv/100 (influx), which is an overestimate since the river discharge has not been accounted. Around 75% of the river discharge in the Mediterranean (0.95 Sv/100) is produced eastward the strait of Sicily [*Struglia et al.*, 2004], then, around 0.7 Sv/100 is discharged in the eastern Mediterranean. Therefore, the mean value of the Strait of Gibraltar is better estimated as 3.7 Sv/100. Thus about two thirds of the Gibraltar influx water continues through the Strait of Sicily.

4. Discussions and Conclusions

[32] Two new improvements in the GRACE TVG data motivate the revisit of the relation among the observed, steric, and mass-induced SLV terms for the Mediterranean Sea: the development of new filters for processing GRACE data, and the fact that the GRACE time series is now 3.5 times longer than in previous studies, leading to a better estimate of seasonal and long-term variations. The total SLV estimate has also been improved with the reprocessing of altimetry data. In particular, the reprocessing has improved the quality of the measurements near the coasts, which is important in a semienclosed basin such as the Mediterranean Sea.

[33] All the SLV terms show clear annual signals with different amplitudes and phases, of which the observed, steric, and mass-induced SLV peak in early October, mid September, and early January, respectively. Comparing these terms, we find a closure of the annual signal of the observed SLV as the net sum of the steric and mass-induced SLV. The observed SLV is mainly driven by steric SLV, which is partially counteracted by the mass-induced SLV. The latter accounts around 1/3 of the steric signal. On the other hand, the OBP from OGCMs is in good agreement with the OBP deduced from GRACE data. Once the latter is converted into mass-induced SLV, the steric term is estimated as the residual "altimetry minus GRACE." The latter agrees in phase with other estimates of steric SLV, although those from ECCO and EN3 data show a lower amplitude. The steric SLV from ECCO accounts 62% of the same signal from Mercator, which may be due to the lack of resolution to fully resolve the eddy energy spectrum, as proposed by *Fu and Smith* [1996] who found an underestimate of its sea level variance by a factor of 2. The Mercator model, which steric SLV estimate is in agreement (within the error estimate) with the "altimetry minus GRACE" and

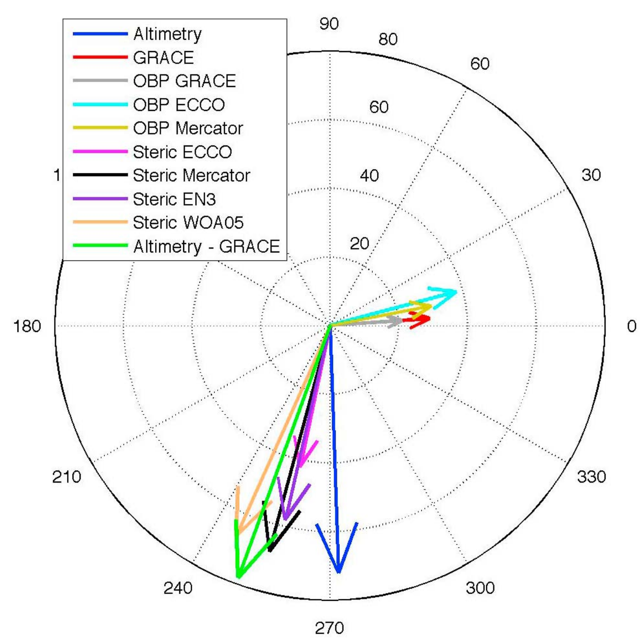

Figure 5. Annual amplitude and phase (equation (4) and Table 1) of different data sets averaged over the Mediterranean Sea. Only "GRACE" refers to mass-induced SLV.

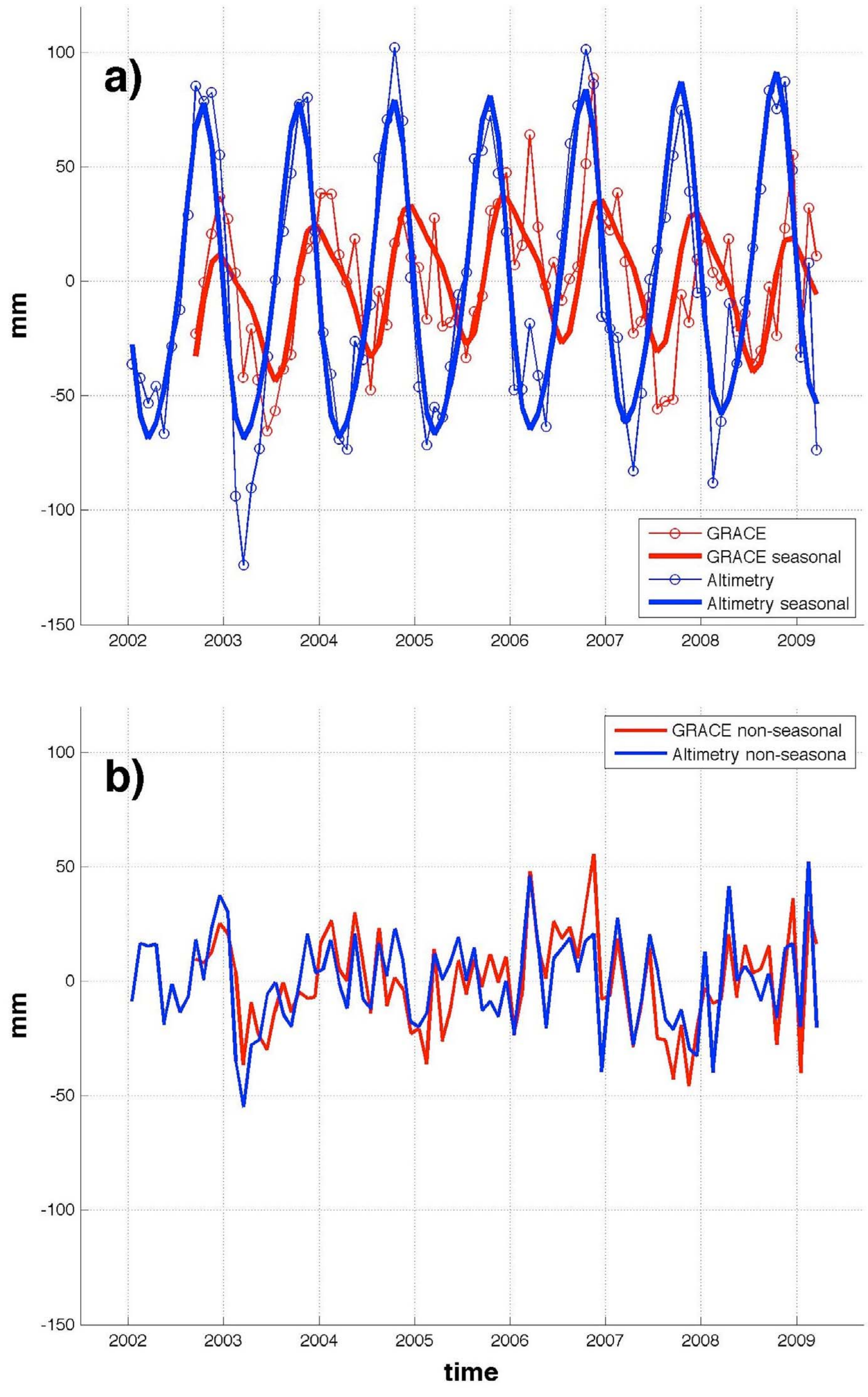

Figure 6. (a) Mean time series for the Mediterranean Sea, excluding Adriatic and Aegean seas, for GRACE as in Figure 4 (thin red line), altimetry (thin blue line), and the corresponding seasonal fits according to equation (4) (thick lines). (b) Nonseasonal residual between the data and the seasonal fit. Red line is the GRACE residual, and blue line is the altimetry residual.

WOA05 signals, shows a bigger amplitude probably because of its higher resolution (0.25 degree) in the target area. However, the lack of full resolution for the eddy energy spectrum in the models cannot explain the low amplitude of the steric SLV from EN3, since the latter come from observations and should recover, at least partially, the process. In the case of the EN3 data, the low amplitude may be due to: (1) The lack of measurements below 500 m depth. (2) A bias due to the spatial distribution of the profiles used to optimally interpolate the grids (there is an average of 185 profiles down to 500 m per month, with a standard deviation of 137), since the annual signal is not uniform in the Mediterranean [e.g., *García et al.*, 2006]. However, the first hypothesis is not very realistic since the steric SLV are

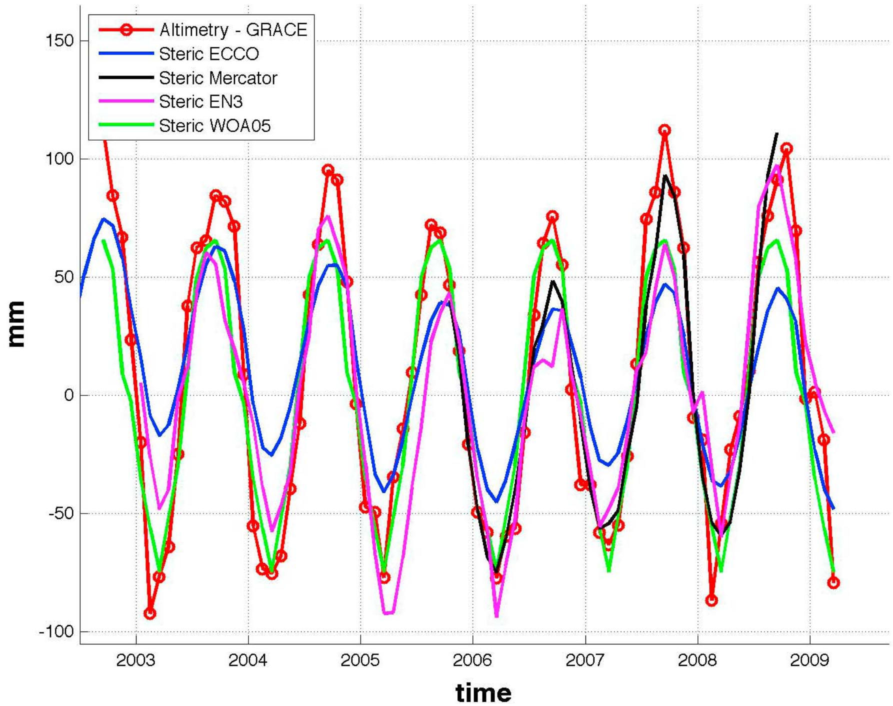

Figure 7. Steric SLV estimated from altimetry minus GRACE mass-induced SLV from Figure 6a (thin blue and red lines) and from salinity and temperature profiles from ECCO (blue line), Mercator (black line), EN3 (magenta line), and WOA05 (green line).

mainly due to temperature changes in the first few hundreds meters from the sea surface. In the second case, we have estimated the averaged time series of all data sets related to steric SLV (altimetry; steric from ECCO, Mercator and WOA05) only using the grid points closest to the profiles for each month. Only SLV amplitudes from altimetry and steric SLV from Mercator decrease by about 10%, but no variations were observed in steric SLV from ECCO or WOA05 data. So, the second hypothesis may only explain a portion of the discrepancies. On the other hand, although the climatological year from historical salinity and temperature records from WOA05 data sets agrees with "altimetry minus GRACE" within the error estimate, it also shows a lower amplitude. However, this discrepancy is not surprising because the annual amplitude in the Mediterranean sea level has been changing during the second half of the twentieth century as derived from tide gauge data, reaching in some stations variations from 30 to 70 mm [*Marcos and Tsimplis*, 2007]. In any case, an overestimate of "altimetry minus GRACE" has to be accepted as cause of the discrepancies.

[34] If the discrepancies cannot be completely explained, what is the correct estimate of the steric SLV annual signal? The OBP and the mass-induced SLV have been estimated from GRACE data, showing that both signals present the same annual phase. However, the OBP from models, which agree with OBP from GRACE within the error estimate, does not agree with the mass-induced SLV estimated from ECCO model as "altimetry minus steric," unlike when estimated from Mercator (see Table 1). The phase disagreement in ECCO disappears when its steric SLV amplitude is multiplied by the factor 1.65. As far as we trust in altimetry, which is a mature technique, the low amplitude of the steric signal seems to be the cause of the disagreement in phase between OBP and mass-induced SLV from ECCO model. In addition, we have observed that the SLV from models do not exactly agree with the altimetry measurements. For example, in the case of ECCO, 20 days shift ahead the altimetry measurements is observed. On the other hand, the "altimetry minus GRACE" signal is based on homogeneous and temporal samplings and can measure variations down to the sea bottom, unlike steric SLV from EN3. Therefore, although inconclusively, we think that the "altimetry minus GRACE" estimate is the most reliable.

[35] Seasonal SLV in the Mediterranean is produced by both steric and mass-induced terms, unlike the intra-annual signal. The seasonal term in mass-induced SLV is related to seasonal water mass fluxes in the basin, mainly precipitation, evaporation, and water exchange through the Gibraltar Strait. The intra-annual SLV is related to mass exchange through the same strait, but driven by winds around it [*Fukumori et al.*, 2007].

[36] The observed SLV does not show any prominent interannual variations, unlike the steric and mass-induced SLV. The latter shows a positive trend up to 2006 and negative trend afterward, while the steric SLV shows an opposite variation, i.e., a negative trend up to 2006 and positive after that.

[37] The water mass budget of the Mediterranean Sea varies annually. The reason is that evaporation exceeds precipitation throughout the year, and the deficit is balanced by water influx from the Atlantic through the Gibraltar Strait. However, this balance is not produced instantaneously, giving rise to the seasonal variations in the water mass budget. The latter has been estimated from GRACE

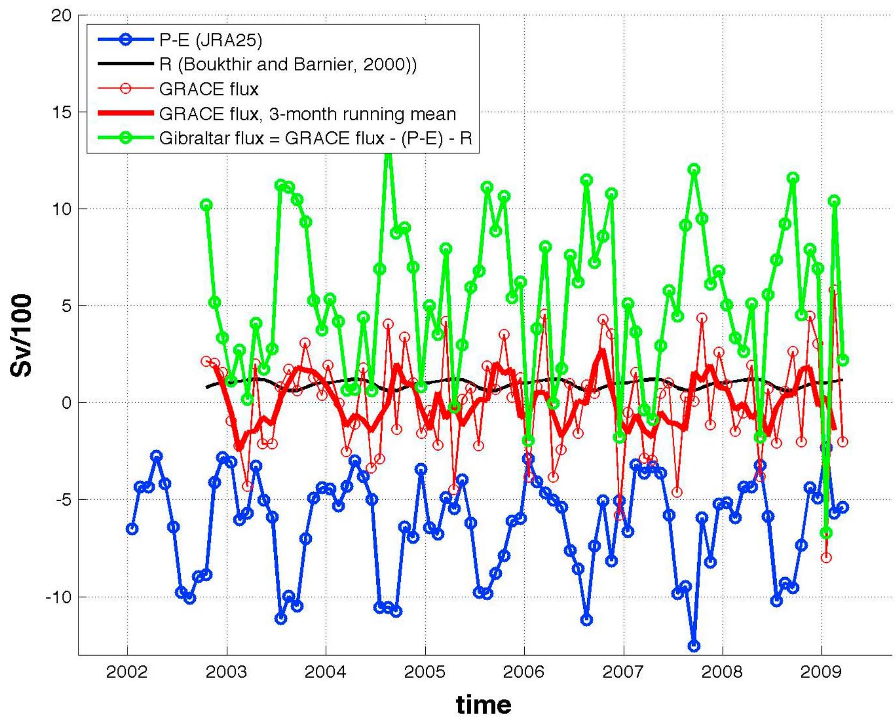

Figure 8. Water mass fluxes averaged over the Mediterranean Sea: total mass flux (F in equation (5)) estimated as month-to-month derivative of the water mass budget estimated from GRACE in Figure 6a (red line), precipitation minus evaporation flux (blue line), climatological river discharge from *Boukthir and Barnier* [2000] (black line), and Gibraltar flux estimated as the residual (green line).

data and the net evaporation from models and observations, via which the water flux through the Gibraltar Strait has been inferred. This flux shows an annual signal of 3.9 ± 0.8 Sv/100 of amplitude and 253° ± 13° of phase (peak in early September), and a mean value of 5.5 Sv/100. Note that the reported errors are formal errors accordingly to equation (4). However, real errors should be bigger due to errors in GRACE data, the atmospheric model, the use of a climatology for the river discharge, the exclusion of the Aegean and Adriatic seas, and neglecting of the water exchange between the Mediterranean and the Black Sea. In any case, the obtained Gibraltar flux signal is between previous reported estimations. Previous similar studies, also based in GRACE data, obtained an annual signal of (60 ± 16 mm/month or 5.4 ± 1.4 Sv/100, 269° ± 13°) [*Fenoglio-Marc et al.*, 2006, 2007], and (18 ± 16 mm/month or 1.6 ± 1.4 Sv/100, 215° ± 62°) [*García et al.*, 2006]. An alternative study based on current meter observations between 1995 and 1998 estimated an annual signal of (7.7 ± 4.4 Sv/100, 234° ± 33°) [*García-Lafuente et al.*, 2002]. A different approach based on tide gauges, sea surface temperature, climatological profiles of temperature and salinity, and OGCM estimated an annual signal of (3.2 ± 2.0 Sv/100, 244° ± 35°) [*García-Lafuente et al.*, 2004]. Thus, our present estimate is between the previous reported values, and agrees very well with the latter. Note that a difference in phase of up to one month exists among the estimates; the reason is not clear as the geophysical process involved in the seasonal variations of the water flux through the Gibraltar Strait is difficult to model [*Gomis et al.*, 2006; *Fukumori et al.*, 2007].

[38] Similar to the estimate of the Gibraltar flux, the net water mass flux through the Strait of Sicily has also been estimated, showing, with respect to the Gibraltar flux, an annual amplitude 40% smaller, a delay around 14 days in

Table 2. Annual Amplitudes and Phases (Indicating the Peak Time in the Year) and Mean Values for the Period 2003–2008 of the Estimated Water Fluxes[a]

	Annual Amplitude (Sv/100)	Annual Phase (°)	Annual Peak Around	Mean (Sv/100)
GRACE flux (F)	1.4 ± 0.8	290 ± 34	21 October	0
GRACE flux 3 month running mean	1.3 ± 0.2	289 ± 11	20 October	0
$P - E$	2.7 ± 0.4	56 ± 8	26 February	−6.3
Gibraltar flux ($G = F - P + E$)	3.7 ± 0.9	253 ± 14	14 September	6.4
Gibraltar flux ($G = F - P + E - R$)	3.9 ± 0.8	253 ± 13	14 September	5.5
Sicilian flux	2.3 ± 0.5	267 ± 12	28 September	3.7

[a]Positive values of both Gibraltar and Sicilian fluxes represent eastward fluxes. The annual amplitude and the mean are multiplied by 100 for clarity, and the units are Sv/100.

the annual signal, and 2/3 of the mean flux. The estimate of both fluxes may be important to constrain local ocean models in Mediterranean subbasins.

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