On the steric and mass-induced contributions to the annual sea level variations in the Mediterranean Sea

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Received 15 March 2005; revised 22 July 2005; accepted 31 January 2006; published 26 September 2006.

[1] The sea level variation (SLV_{total}) is the sum of two major contributions: steric and mass-induced. The steric SLV_{steric} is that resulting from the thermal and salinity changes in a given water column. It only involves volume change, hence has no gravitational effect. The mass-induced SLV_{mass}, on the other hand, arises from adding or subtracting water mass to or from the water column and has direct gravitational signature. We examine the closure of the seasonal SLV budget and estimate the relative importance of the two contributions in the Mediterranean Sea as a function of time. We use ocean altimetry data (from TOPEX/Poseidon, Jason 1, ERS, and ENVISAT missions) to estimate SLV_{total}, temperature, and salinity data (from the Estimating the Circulation and Climate of the Ocean ocean model) to estimate SLV_{steric}, and time variable gravity data (from Gravity Recovery and Climate Experiment (GRACE) Project, April 2002 to July 2004) to estimate SLV_{mass}. We find that the annual cycle of SLV_{total} in the Mediterranean is mainly driven by SLV_{steric} but moderately offset by SLV_{mass}. The agreement between the seasonal SLV_{mass} estimations from $SLV_{\text{total}} - SLV_{\text{steric}}$ and from GRACE is quite remarkable; the annual cycle reaches the maximum value in mid-February, almost half a cycle later than SLV_{total} or SLV_{steric}, which peak by mid-October and mid-September, respectively. Thus, when sea level is rising (falling), the Mediterranean Sea is actually losing (gaining) mass. Furthermore, as SLV_{mass} is balanced by vertical (precipitation minus evaporation, $P-E$) and horizontal (exchange of water with the Atlantic, Black Sea, and river runoff) mass fluxes, we compared it with the $P-E$ determined from meteorological data to estimate the annual cycle of the horizontal flux.

Citation: García, D., B. F. Chao, J. Del Río, I. Vigo, and J. García-Lafuente (2006), On the steric and mass-induced contributions to the annual sea level variations in the Mediterranean Sea, *J. Geophys. Res.*, *111*, C09030, doi:10.1029/2005JC002956.

1. Introduction

[2] The absolute total sea level variation (SLV) consists of two contributions: steric and mass-induced:

$$SLV_{\text{total}} = SLV_{\text{steric}} + SLV_{\text{mass}} \tag{1}$$

Here the SLV is defined as the temporal anomaly as a function of geographical location relative to the "static" mean geoid and in reference to the terrestrial reference frame of the Earth. We consider timescales longer than monthly; on such timescales, the SLV is a critical indicator of the oceanographic and climatic processes. In that regards it is desirable to know the exact share, or relative importance, of the two SLV contributions as a function of location and time. Such knowledge will lead to valuable understanding about ocean dynamics as well as global climatic changes.

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0148-0227/06/2005JC002956

[3] SLV_{steric} results from the volumetric expansion or contraction induced by variations of water temperature T and salinity S in the water column, basically a baroclinic phenomenon. SLV_{mass}, on the other hand, is simply a result of the addition (e.g., precipitation, river runoff, melting of land ice) or subtraction (e.g., evaporation, dam impoundment on land) of water mass to or from the water column, basically a barotropic effect. The relative share of the two SLV contributions is a strong function of location and time. For example, subject to large temperature swings in the course of the year, middle and low latitudes often see relatively larger steric contribution in the SLV_{total}. At higher latitudes and locations where dynamic variability is strong, the mass-induced effect can become the major contibutor in SLV_{total}. In general, at any given location the two contributions superimpose at their own seasonal phasing; they can augment or oppose each other depending on meteorology and ocean dynamics. There are also large nonseasonal phenomena that manifest both in steric and mass-induced SLV, such as ENSO and Pacific Decadal Oscillation in the Pacific, and the North Atlantic Oscillation. Both SLV contributions presumably have long-term trends as well because of climate changes that result in global warming and melting of land ices.

[4] $SLV_{\rm total}$ has been monitored by radar altimetry from the vantage point of artificial satellites for over two decades now. In particular, TOPEX/Poseidon (T/P) has acquired continuous and near-global coverage of $SLV_{\rm total}$ at a precision of 2–3 cm since 1992, and its successor Jason 1 is now achieving comparable data, if not of even higher precision [*Luthcke et al.*, 2003]. $SLV_{\rm steric}$ can be estimated using the equation of state for seawater when T and S observations are available, typically through in situ CTD and XBT casts collected from ships during oceanographic surveys, as well as the Argo floats (for detail information see ⟨www.argo.ucsd.edu⟩). At seasonal timescales, the steric effect is felt mostly in the upper layer of the ocean. $SLV_{\rm steric}$ only involves volume changes of the same amount of water, and hence has virtually zero gravitational signature.

[5] On the other hand, $SLV_{\rm mass}$ is accompanied by direct gravitational signature according to its temporal and spatial variations, and hence can be observed gravitationally. Since its launch in March 2002, the NASA/DLR dual-satellite space mission Gravity Recovery and Climate Experiment (GRACE) has mapped the time variable gravity (TVG) signals globally with a monthly temporal resolution and a spatial resolution of about 1500 km [*Tapley et al.*, 2004; *Wahr et al.*, 2004]. From the TVG one can readily and uniquely determine the mass variation that produce the TVG signal, assuming it comes from the Earth's surface [e.g., *Wahr et al.*, 1998; *Chao*, 2005]. The $SLV_{\rm mass}$ follows immediately from the surface mass variation (and the known density of seawater).

[6] Ideally, if we had perfect determination of $SLV_{\rm total}$ (e.g., from satellite altimetry), $SLV_{\rm steric}$ (e.g., from in situ measurements), and $SLV_{\rm mass}$ (e.g., from TVG measurement), they should obey the simple equation (1). Alternatively, then, knowing any two quantities perfectly, the third can be determined. The reality is of course far from the case. The data involved are noisy, contaminated, and incomplete (if not sparse) under limited sampling in time and space. Conceivably, the use of an ocean general circulation model (GCM) can serve as a tool to enforce equation (1) by assimilating available data of all three data types into a theoretical framework based on physics. In fact, in this paper we make use of the $SLV_{\rm steric}$ estimated from such a GCM, namely the Estimating the Circulation and Climate of the Ocean (ECCO) ocean model [*Lee and Fukumori*, 2003].

[7] The target subject of this paper is the Mediterranean Sea. A semienclosed basin, the Mediterranean Sea exchanges water with the open sea only through the narrow Strait of Gibraltar, with the Black Sea through the even narrower Dardanelles and Bosphorus Straits, and with the land by the moderate runoff of rivers such as the Nile. If the fluxes through these waterways are closely monitored or modeled, then the only other water exchange is with the atmosphere in the form of precipitation minus evaporation (P–E), which can be modeled given sufficient meteorological data. Thus the Mediterranean water budget can in principle be quite tractable. In this paper we will examine a host of direct and indirect measurement data to estimate the various SLV in the Mediterranean with respect to equation (1). Because of the short time span of the GRACE TVG data (for estimating $SLV_{\rm mass}$), we will concentrate on the seasonal variability, which is dominant in all observables.

[8] To first order, it is not clear to what extent the total Mediterranean water budget undergoes a seasonal cycle [*García-Lafuente et al.*, 2002; *Bouzinac et al.*, 2003]. Such a cycle implies an out-of-phase exchange of water through the air-sea interface ($P-E$) and through the Straits and rivers. In the long term, the latter compensates the fresh water deficit produced by the former in order to preserve the water mass of the Mediterranean, but the compensation does not have to happen on a short-term basis, hence a seasonal signal of the mass budget in the Sea. Up to now, the seasonal mass imbalance at basin scale has been estimated in indirect ways; these data had been too noisy to elucidate on the weak signals at hand [*García-Lafuente et al.*, 2002; *Bouzinac et al.*, 2003; *Larnicol et al.*, 1995].

2. Data and Processing

[9] For $SLV_{\rm total}$, we use a combined monthly solution from T/P [*Fu and Cazenave*, 2001], Jason 1, ERS and ENVISAT altimetry missions, on a $1^\circ \times 1^\circ$ regular grid. There are 310 such grid points in the Mediterranean Sea. The time span is January 1993 to July 2004. All standard corrections were applied to the altimetry data, including the inverted barometer effect applied to reduce aliasing errors, although it may introduce slight errors of its own by violating water mass conservation in the semienclosed sea.

[10] To estimate $SLV_{\rm steric}$, the temperature T and salinity S fields from two ECCO ocean model products are used: the JPL simulation and the JPL adjoint smoothed wind-driven (http://www.ecco-group.org). The simulation uses the NCEP reanalysis as forcing except their time means were replaced with those of Comprehensive Ocean-Atmosphere Data Set (COADS) (for details, see *Lee and Fukumori* [2003]). The smoothed wind-driven run is based on a correction to this NCEP wind forcing estimated by assimilating altimetry data from T/P and Jason 1, and temperature profiles available on Global Telecommunication System (GTS from in situ measurements by XBT, CTD, and Argo floats). Other components of the forcing are the same as the simulation (i.e., NCEP plus COADS correction). Although both simulation and assimilation products have been processed, only the former is shown here as both produce nearly exactly the same results. Data profiles are from surface to the (nonuniform) sea bottom at each point on a $1^\circ \times 1^\circ$ regular grid, and the time span used is 1997–2004 with a time step of ten days. The state law algorithm for computing the corresponding $SLV_{\rm steric}$ is adopted from *Pond and Pickard* [1986].

[11] To estimate $SLV_{\rm mass}$, we use the 22 monthly sets of normalized spherical harmonic Stokes coefficients provided by the GRACE Project, for the period April 2002 to July 2004. The data include Stokes coefficients C_{lm} and S_{lm} up to degree 120 but among them only harmonic solutions up to degree $\sim$15 are sufficiently well determined for TVG [*Tapley et al.*, 2004; *Wahr et al.*, 2004], corresponding to a spatial resolution of about 1500 km (see below). The monthly spherical harmonic TVG field is readily converted into surface mass variations by:

$$\Delta\sigma(\theta,\lambda) = \frac{a\rho_E}{3}\sum_{l=0}^{\infty}\sum_{m=0}^{l}\frac{(2l+1)}{(1+k_l')}P_{lm}(\cos\theta) \cdot [\Delta C_{lm}\cos m\lambda + \Delta S_{lm}\sin m\lambda] \qquad (2)$$

assuming the latter only occur on the Earth surface [*Wahr et al.*, 1998; *Chao*, 2005], where (θ, λ) = (colatitude, longitude), $\Delta\sigma$ is the estimated mass variation, ρ_E is the mean density of the Earth, k'_l is the lth load potential Love number, P_{lm} is the 4π normalized associated Legendre function of degree l and order m, and ΔC_{lm} and ΔS_{lm} are the time variable Stokes coefficients. Mass variations are further converted into ''water thickness equivalent'' (WTE) in units of mm by dividing with the water density assumed to be 1000 kg/m^3 (thus each kg/m^2 is equivalent to 1 mm WTE). Note that, however, a factor of 1/1.029 should be applied to the WTE to convert it into the observed SLV owing to the ocean water density of 1029 kg/m^3.

[12] A mass variation in the water column produces a deformation in the sea bottom. As a consequence, a mass redistribution arises producing a variation of the potential. In equation (2), the division by the factor $(1 + k'_l)$ undoes that variation of the potential and hence the original mass variation is recovered. Furthermore, when SLV_{mass} is indirectly estimated from equation (1), the deformation produced by mass change is implicitly observed. For that reason, to fairly compare with SLV_{mass} estimated through equation (2), the sea bottom deformation should be corrected, which can be estimated by:

$$\Delta H(\theta,\lambda) = a\sum_{l=0}^{\infty}\sum_{m=0}^{l}\frac{h'_l}{(1+k'_l)}P_{lm}(\cos\theta) \cdot [\Delta C_{lm}\cos m\lambda + \Delta S_{lm}\sin m\lambda] \tag{3}$$

where h'_l is the lth load deformation Love number [*Munk and Macdonald*, 1960]. This effect proves to be negligible numerically, although important from a theoretical point of view and adopted in our computations.

[13] The GRACE TVG data have beforehand been corrected for the following: the atmospheric effect according to the ECMWF GCM output, the short-period oceanic effect based on a barotropic ocean GCM [*Flechtner*, 2003], the solid Earth tides (including solid pole tide), ocean tides (including the ocean pole tide as a consequence of the solid pole tide via an equilibrium response [*Wahr*, 1985; *AVISO*, 1996], but not including the effects of loading and self-gravitation of the ocean pole tide [*Desai*, 2002]), as well as the routine satellite orbit perturbations of secular polar motion, N body and general relativistic effects. In order to compare between GRACE TVG and altimetry data, these differences should be considered, as have been in previous studies [e.g., *Chambers et al.*, 2004].

[14] Thus we first add back the barotropic ocean GCM effect to the GRACE TVG field. Secondly, while zero in GRACE data (because the center of mass of the Earth is taken to be the origin of the reference frame), the degree 1 terms are nonzero in altimetry data. Thus we add to the GRACE TVG field the estimated degree 1 terms calculated from an estimation of the annual and semiannual motion of the mass center of the Earth according to *Chen et al.* [1999]. Thirdly, the degree 2 order 1 GRACE TVG coefficients have been corrected for secular variations but not corrected in altimetry data. Therefore we add back the secular variations to those coefficients. Besides, as the degree 2 order 0 coefficient,, which represents Earth's oblateness [*Cox and Chao*, 2002], is not well observed by GRACE [*Tapley et al.*, 2004], we have replaced it for a more accurate time series estimated from satellite laser ranging (SLR) (C. Cox, personal communication, 2005). In order to remove the atmospheric effect in the latter, we subtract the degree 2 order 0 coefficients of the atmospheric correction applied to GRACE data [*Flechtner*, 2003]. Finally, the only disturbing potential related to the polar motion assumed in altimetry correction is that produced by the solid earth pole tide. The disturbing potential produced by the latter leads to an equilibrium response of the sea level, called ocean pole tide, which is corrected in altimetry [*AVISO*, 1996]. However, the water mass displaced by that response produces extra effects of loading and self-gravitation, which are not corrected in altimetry [*Desai*, 2002]. In GRACE data, the solid earth pole tide is corrected, so no ocean response is observed because of the related disturbing potential, but the effects of loading and self-gravitation of the ocean pole tide are not corrected. Therefore altimetry and GRACE partially correct the ocean pole tide in the same way and no inconsistency arises when comparing both data sets.

[15] Ancillary sources of data are processed in order to examine the water mass exchange fluxes in the Mediterranean Sea. In particular, we shall use the monthly $P-E$ field derived from the proxy atmospheric humidity data as provided by the NCEP atmospheric GCM (B. F. Chao and A. Y. Au, Global hydrological budget derived from atmospheric circulation model and GRACE time-variable gravity data, submitted to *Earth Interactions*, 2005, hereinafter referred to as Chao and Au, submitted manuscript, 2005), on a 2.5° × 2.5° grid for the period January 2002 to July 2004. These data are later combined with GRACE TVG data to estimate the water mass exchange through the Gibraltar Strait neglecting the moderate river runoffs. This result will be compared with historical data of the net Gibraltar flux of water, as estimated from current meter observations taken during the CANIGO project between October 1995 and May 1998 [*García-Lafuente et al.*, 2002].

3. Results and Discussions

3.1. Different Contributions to the Annual Mediterranean Sea Level Signal

[16] We shall first examine the spatial distribution of the seasonal signals. The annual signal can be extracted by least squares fitting a given signal with seasonal sinusoids in terms of amplitude (A) and phase (φ):

$$signal = A\cos(\omega_a t - \varphi) \tag{4}$$

where ω_a is the annual frequency and t denotes the time in months. Note that by this definition the phase φ corresponds to the time of the maximum positive amplitude during the year. An alternative form of equation (4) that includes an additional semiannual harmonic has also been examined, which shows that semiannual amplitudes are an order of magnitude smaller than annual signals while the latter remain virtually unmodified. Thus, for the present purpose, equation (4) is adequate to study the seasonal cycle and is used throughout this paper. Figures 1a and 1b show the resultant annual A and φ estimates for SLV_{total} (altimetry) and SLV_{steric} (ECCO ocean GCM), respectively. The studied period depends on the availability of each data set.

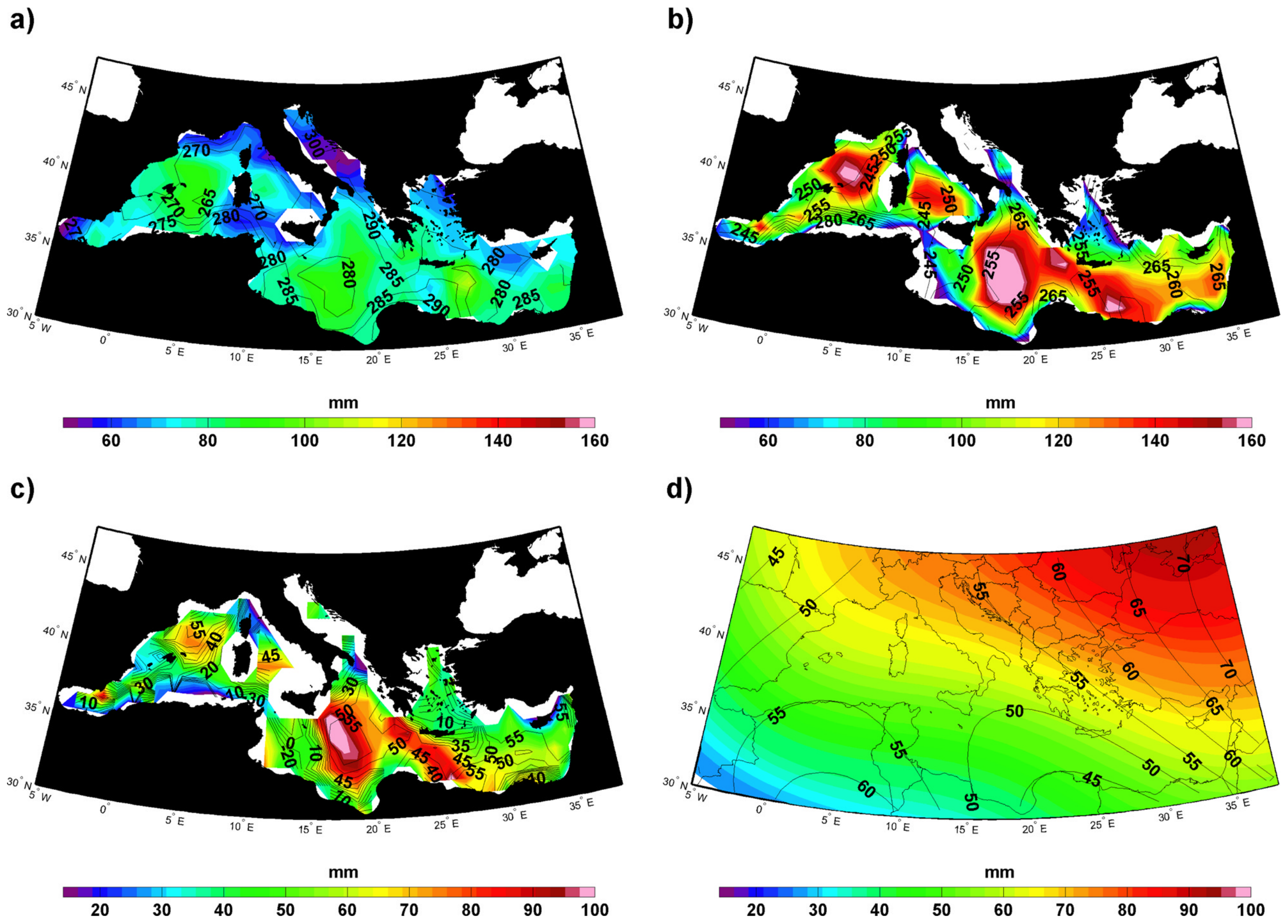

Figure 1. Annual amplitude A (color scale in mm) and phase φ (contour lines in degrees) from equation (4) for different data sets: (a) SLV_{total} from altimetry; (b) SLV_{steric} from ECCO model; (c) SLV_{mass} = (Figure 1a)–(Figure 1b); (d) SLV_{mass} from GRACE data.

[17] Figure 1a shows a SLV_{total} annual amplitude of roughly 60–80 mm except for the Adriatic Sea (with ~50 mm) and four regions (the southern Levantine Basin, the Tyrrhenian Sea, the Ionian Sea and the central part of the Western Mediterranean Basin) where values up to 80–100 mm are reached. This spatial pattern is similar to Figure 1b that shows the annual amplitude of SLV_{steric} with values of ~90 mm, except for the same four regions of enhanced amplitude as in Figure 1a where values around 130–160 mm are observed. Besides, low amplitudes less than 60 mm are observed in the Adriatic and Aegean seas and near the Tunisian coast. The maximum positive amplitude of SLV_{total} occurs between 270° and 285° (first half of October) in the whole basin and ~300° in the Adriatic Sea (Figure 1a), while that of SLV_{steric} takes place around one month earlier (values between 240° and 260°, that is, the first half of September). Both of them are, not surprisingly, rather homogeneous. A noteworthy feature in these figures is the west-to-east propagation of both sea level signals, with the Western Mediterranean Basin leading the Eastern Basin by 15 days (10° to 20°) or so.

[18] From these figures, it is clear that SLV_{total} and SLV_{steric} do not match each other. Not only SLV_{steric} has greater amplitude on average, but its phase also leads that of SLV_{total} by around 30°. Their difference, $SLV_{\mathrm{total}} - SLV_{\mathrm{steric}}$, which is an indirect estimate of SLV_{mass} according to equation (1), is clearly nonvanishing, as shown in Figure 1c. Its annual amplitude is 30–60 mm, with two localized regions showing more than 90 mm, and its annual phase is between 10° and 55° (mid-January and late February), except for a localized region in the Western Basin with a phase of 330° (or −30°). Now compare it with Figure 1d, which gives SLV_{mass} estimated from GRACE. Its annual amplitude is ~50 mm and its phase range from 45° to 65° (second half of February), which propagates northeastward in the Levantine Basin and is quite homogeneous in the Western Basin. Aside from the much lower spatial resolution of the GRACE map, which does not allow the detection of the small features observed in figure 1c, the agreement between both approaches is reasonably good in general (or more precisely in average), considering that (1) they are completely independent data types with uncorrelated noises and (2) Figure 1c is a residual signal between two large varying fields. Particularly notable is the large phase difference of SLV_{mass} with SLV_{total} or SLV_{steric}. We conclude that the three data sets tell a consistent story as far as the annual variability is concerned.

[19] To proceed, we examine only the mean annual signal averaged over the entire Mediterranean Sea region. We do so because all data sets are to be studied in conjunction with

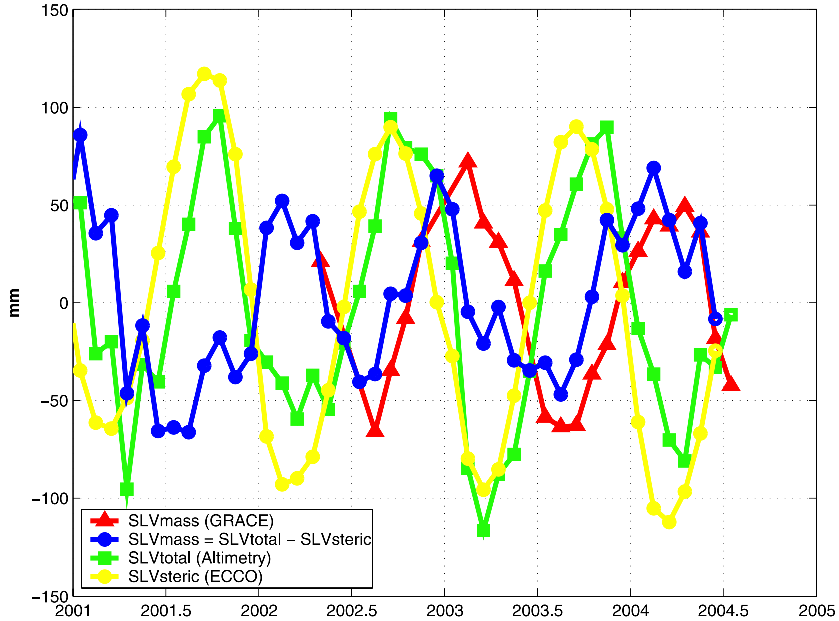

Figure 2. The time series represents the monthly mean values over the Mediterranean Sea for several data sets. Red curve: SLV_{mass} from GRACE data; green curve: SLV_{total} from altimetry data; yellow curve: SLV_{steric} from ECCO assimilation model; blue curve: SLV_{mass} estimated according to equation (1), $SLV_{\text{mass}} = SLV_{\text{total}} - SLV_{\text{steric}}$.

GRACE data and GRACE can barely resolve the detail within the Mediterranean Sea because of its relatively low spatial resolution. Thus we average both altimetry and steric height data in the Mediterranean Sea to obtain the mean time series variations (Figure 2).

[20] In order to isolate the Mediterranean area to calculate its mean TVG signal from the GRACE data and to minimize the spectral leakage and data error for higher degrees, a Gaussian filter as described by *Swenson and Wahr* [2002] is applied to obtain the estimated mean mass anomaly $\Delta\tilde{\sigma}_{Med}$ in the Mediterranean Sea as follows:

$$\Delta\tilde{\sigma}_{\text{Med}} = \frac{1}{\Omega_{\text{Med}}} \int_{\text{Med}} \Delta\sigma(\theta,\phi)\bar{W}(\theta,\phi)d\Omega \qquad (5)$$

where Ω denotes solid angle, $\bar{W}(\theta, \phi)$ mimics value 1 over the Mediterranean Sea and 0 otherwise, changing smoothly (taken as 1000 km) at the boundary and truncating $\Delta\sigma(\theta, \phi)$ at degree 15 (see section 2). For each month of data this leads to a single $\Delta\tilde{\sigma}_{\text{Med}}$ value obtained by weighted sum of all the coefficients up to degree 15. The end product, when converted into WTE, is a monthly time series for mean SLV_{mass}.

[21] Figure 2 presents all the monthly mean time series over the Mediterranean Sea, while Table 1 shows the numerical estimates of the spatially averaged (A, φ) from equation (4). The uncertainty estimates are those of the 95% formal error during the least squares fit procedure.

[22] Figure 2 and Table 1 show the same good agreement between SLV_{total} and SLV_{steric} observed in Figure 1 in both annual amplitude and phase. We notice (1) the amplitude of SLV_{total} is ~10 mm lower and peaks ~23 days later than SLV_{steric}, (2) the SLV_{mass} estimated from GRACE shows an annual amplitude of 55 mm and a phase of 52° (mid-February) which are noticeably different from those of SLV_{total} and SLV_{steric}, and (3) the estimation of SLV_{mass} as $SLV_{\text{total}}-SLV_{\text{steric}}$ from equation (1) agrees very well with GRACE-observed SLV_{mass}. Thus the annual SLV_{mass} in the Mediterranean is ~230° (8 months) ahead or ~130° (4 months) lagging with respect to SLV_{total}. The similitude of SLV_{steric} and SLV_{total} curves in Figure 2 indicates clearly that Mediterranean sea level is mainly driven by the steric changes, while SLV_{mass} amounts to about one third of SLV_{total} but having a quite different phase. For the first time it is shown how annual steric and mass-induced changes in the Mediterranean Sea counteract each other to produce the net sea level variation.

[23] We have experimented with the same scheme but breaking down the Mediterranean Sea into three regions: (1) Western Basin, (2) Central Mediterranean (including Ionian Sea) and Adriatic Sea, and (3) Levantine Basin. The individual mean SLV_{mass} (not shown) show that the annual signals are quite similar in all regions, and almost identical to that of the whole Mediterranean in Figure 2. This is consistent with the fact that the annual SLV_{mass} is homogeneous in the whole basin (see Figure 1) with little exchange

Table 1. Annual Amplitude (A) and Phase (φ) of the Different Spatially Averaged Monthly Time Series and the Period Covered by the Data Sets[a]

	Period, month/year	A, mm	φ
GRACE	04/02–07/04	24 ± 14	50° ± 31°
Alt − ECCO steric	04/02–07/04	38 ± 16	16° ± 27°
Altimetry	04/02–07/04	83 ± 13	281° ± 10°
ECCO steric	04/02–07/04	94 ± 5	258° ± 3°
$P-E$ (NCEP)	04/02–07/04	31 ± 8/month	7° ± 17°
F (equation (5))	09/02–07/04	18 ± 16/month	215° ± 62°

[a]The uncertainty indicates 95% formal error.

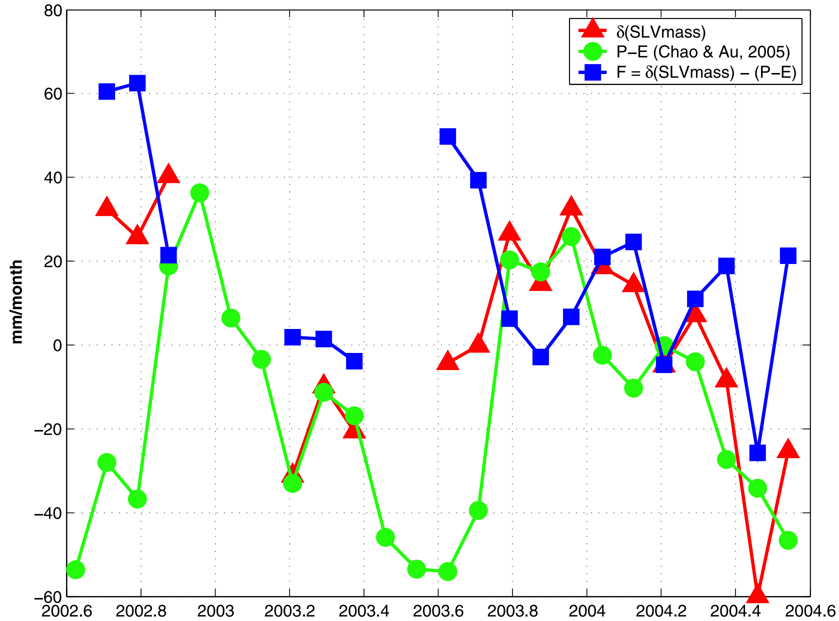

Figure 3. The time series represents the monthly mean values over the Mediterranean Sea for several data sets. Red curve: $\delta(SLV_{\mathrm{mass}})$ from GRACE data; green curve: $P - E$ field from Chao and Au (submitted manuscript, 2005); blue curve: water mass flux F from the difference between the above two curves according to equation (6).

among the regions. However, it is interesting to point out a possible gaining of water in region 3 at the expense of region 1 on an interannual timescale. At present the data span is still too short to tell.

3.2. Mass Signal and the Water Fluxes

[24] Besides the above indirect scheme of determine SLV_{mass} by subtracting SLV_{total} (altimetry) with SLV_{steric} (ocean GCM) according to equation (1) [see also *Bouzinac et al.*, 2003; *Larnicol et al.*, 1995], one can alternatively observe the net barotropic flow through the Strait of Gibraltar from in situ sensors and compare it with the $P - E$ estimates in the area [*García-Lafuente et al.*, 2002]. Conversely, now that SLV_{mass} can be directly measured by GRACE, we can deduce this mass signal arising as the balance between the "horizontal" water mass flux F and the vertical flux $P - E$, taking the form

$$\delta(SLV_{\mathrm{mass}}) = F + (P - E) \qquad (6)$$

where δ indicates the month-to-month incremental change which is calculated from GRACE data. Only 18 monthly values of $\delta(SLV_{\mathrm{mass}})$ have been obtained subject to the availability of GRACE product. They are shown in Figure 3.

[25] On the other hand, the monthly $P - E$ field as estimated by Chao and Au (submitted manuscript, 2005) (see section 2) are averaged over the Mediterranean Sea. The annual signal (A, φ) obtained according to equation (4) is $A = 31$ mm/month and $\varphi = 7°$ (early January) during the period of Figure 3. It is interesting to note that the mean $P - E$ in the Mediterranean is negative in general, with no more than three positive monthly values per year. The mean value is -15.7 mm/month during the time period in Figure 3, amounting to a total of 19 cm/year of net loss of water, which is considerably less severe than the historically reported values [*Bryden et al.*, 1994; *Boukthir and Barnier*, 2000].

[26] An estimate of the water mass flux F follows immediately from equation (6) (see Figure 3). F comes primarily from the flux through the Gibraltar Strait, while the river runoff and the exchange with the Black Sea are negligible in comparison [*Bethoux and Gentili*, 1999]. Its estimated annual signals are $A = 17$ mm/month and $\varphi = 263°$ (late September). The yearly mean value of F cannot be readily estimated using GRACE data because there are only 18 months of δ (SLV_{mass}). Nevertheless, as long as the interannual variability and trends are insignificant, the Mediterranean mean mass content does not vary much from year to year and the mean F should be completely offset by $P - E$ flux.

[27] The F estimate can be compared with historical results as reported by *García-Lafuente et al.* [2002], who estimated the main contributor of F (the flux through the Gibraltar Strait) from measurements of three arrays of current meters situated in the East side of Gibraltar between October 1995 and May 1998. They give an annual influx amplitude of 0.077 ± 0.044 Sv (1 Sv = 10^6 m^3/s), with annual phase $234° \pm 33°$ (late August). For a Mediterranean Sea area of 2.57×10^{12} m^2, this amplitude is equivalent to 78 ± 44 mm/month. Our estimate for F above has a much lower annual amplitude; it is not clear which estimate is more accurate but the different period of time should be noted. However, they agree remarkably well in phase, which is encouraging because of the very different procedure and data sets used.

4. Conclusions

[28] We have demonstrated the combined application of several remote sensing data types to understand the behav-

ior of SLV in the Mediterranean Sea. We have produced monthly time series of the total SLV from altimetry (SLV_{total}), which has an annual amplitude of 83 mm and peaks in mid-October. We have also estimated monthly steric SLV from *T* and *S* outputs of ECCO ocean model (SLV_{steric}), with amplitude of 94 mm and peak phase at mid-September, differing not much from those of SLV_{total}. To close this (secondary) difference calls for a mass signal SLV_{mass}, according to equation (1). The similar behavior between SLV_{total} and SLV_{steric} indicates that the Mediterranean annual SLV_{total} is mainly driven by steric changes and only moderately offset by mass-induced changes. This SLV_{mass} contribution is found to match remarkably well with that detected independently by GRACE, which has an annual amplitude of 55 mm and the peak phase at mid-February, the latter is almost half a cycle later than SLV_{total} or SLV_{steric}. These results confirm previous indirect measurements [*García-Lafuente et al.*, 2002; *Bouzinac et al.*, 2003; *Larnicol et al.*, 1995] which indicated the existence of an annual signal in the Mediterranean mass budget. Overall our results show that, during the annual cycle, when the sea level is rising (falling) in the Mediterranean Sea, it is also losing (gaining) mass. Such a phase difference is not surprising (another recent indication was reported by *Chao et al.* [2003] in the Pacific), as SLV_{steric} and SLV_{mass} are governed by different processes.

[29] Previous studies of global mass variations based on GRACE observations show that the annual amplitude in the open ocean is significantly lower (at 7–9 mm) [*Chambers et al.*, 2004] than the values found here in the Mediterranean Sea (55 mm), while certain land values can be quite larger ($\sim$150 mm in the Amazon River basin and the Bay of Bengal) [*Wahr et al.*, 2004].We note that several studies have revealed that the Mediterranean Sea usually undergoes a different behavior than the global ocean. For example, *Tsimplis and Baker* [2000] suggested a sea level drop in the Mediterranean since 1960s to early 1990s while the global sea level was rising in general [*Douglas*, 1997]. Likewise, *Cazenave et al.* [2001] showed that in the 1990s the rate of sea level rise in the Levantine Basin of the Mediterranean was an order of magnitude higher than the global value, and *Vigo et al.* [2006] reported a change of sea level tendency in some Mediterranean regions which has little relation with the global ocean.

[30] While providing an effective means toward the understanding of the geophysical processes that are responsible for ocean dynamic behaviors, the low spatial resolution of the standard GRACE data product proves to be an impediment to our present study, thereby little TVG detail within the Mediterranean can be revealed. Alternative GRACE data processing approaches aimed at improving spatial and temporal resolutions are currently explored. For example, *Rowlands et al.* [2005] demonstrated a factor of 3 improvements in both temporal and spatial resolutions through a "mascon" analysis of GRACE orbit data. Future studies taking advantage of such higher-resolution data can potentially yield much refined conclusions.

[31] **Acknowledgments.** The authors thank NASA Pathfinder project and Andrew Au for assistance in data processing. The altimeter products were produced by the CLS Space Oceanography Division as part of the Environment and Climate EU ENACT project (EVK2-CT2001-00117) and with support from CNES. GRACE data are provided by the NASA/DLR GRACE Project via the Center for Space Research website in the University of Texas at Austin. The ECCO model is a contribution of the Consortium for Estimating the Circulation and Climate of the Ocean funded by the National Oceanographic Partnership Program. This work, as part of the lead author's Ph.D. dissertation, is supported by the Spanish Science and Technology Ministry Projects ESP2001-4533/PE, ESP2005-02212 and REN2003-01608/MAR, by the Valencian regional government grant ACOMP06-120, and by NASA's Physical Oceanography Program.

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