# Mediterranean Sea level and barotropic flow through the Strait of Gibraltar for the period 1958–2001 and reconstructed since 1659

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[1] Sea level values from a two-dimensional model of the Mediterranean Sea forced by atmospheric pressure and wind are used to estimate the barotropic flow through the Strait of Gibraltar for the period 1958–2001. The Mediterranean mean sea level derived from the model ranges between ±20 cm with a standard deviation of 5.5 cm and is correlated to the North Atlantic Oscillation (NAO) index. Thus NAO historical data and reconstructions are used to derive the Mediterranean Sea level variability from 1659 until 2001. The accuracy of the reconstruction is estimated in 2.7 cm for monthly mean values, 0.41 cm for annual mean values, and 0.22 cm for decadal mean values (0.48 cm for decadal winter mean sea level). The barotropic flow through the strait is computed from the model output as the time derivative of the total volume of the basin. During the period 1958–2001 the estimated daily flow ranges between ±2.7 Sv, with a standard deviation of 0.56 Sv. The dominant periodicities are in between 1 and 2 weeks. At these scales the model successfully reproduces previously published flow estimates based on current meter observations, which confirms that atmospheric pressure and wind dominate the intraseasonal variability of the flow. For the annual cycle the variability of the atmospherically induced flow is similar to the variability of the flow induced by the evaporation-precipitation (E-P) budget (±0.025 Sv), though absolute values of the first are about a third of the latter. At longer timescales the atmospheric contribution is much smaller than the E-P induced flow.

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## 1. Introduction

[2] The Strait of Gibraltar enables and at the same time limits the communication of the Mediterranean Sea to the global ocean. With a minimum depth of around 280 m at the Camarinal sill and a minimum width of around 14 km at Tarifa, the strait permits the balancing of the negative Mediterranean Seawater budget (the evaporation within the basin exceeds precipitation and freshwater inflow from rivers and the Black Sea [*Garrett et al.*, 1993; *Gilman and Garrett*, 1994]). The continuous inflow of Atlantic water is highly modulated by the tidal signal and the atmospheric forcing. Hence *Candela et al.* [1989] explained up to 80% of the observed variability of the currents in the strait on the basis of an almost barotropic mode, which is well correlated to the cross strait sea level component [see, e.g., *Tsimplis and Bryden*, 2000]. The Strait of Gibraltar also acts as a choking point on direct meteorological forcing in the Mediterranean Sea, making the response of sea level in the basin to be underisostatic (that is, smaller in magnitude than the −1 cm per mbar of change of atmospheric pressure) for pressure signals with periods smaller that 15 days [*Garrett and Majaess*, 1984; *Lascaratos and Gacic*, 1990].

[3] Because the knowledge of the subinertial flow through the Strait of Gibraltar enables the knowledge of different key processes, several efforts have been made along the last century to determine the size of the exchanges [*Schott*, 1915; *Bryden et al.*, 1994; *Hopkins*, 1999; *Tsimplis and Bryden*, 2000; *García-Lafuente et al.*, 2002a]. An interesting review of the state of the art and uncertainties about the flow through Gibraltar is given by *Candela* [2001]. Estimates of the Atlantic inflow range between 0.72 and 1.60 Sv (1 Sv = $10^6$ $m^3 s^{-1}$), the outflow ranges between 0.80 and 1.68 Sv and the net exchange is of the order of tenths of Sv. It is not clear to what extent the variability in the estimates reflects changes in the different forcings or it is an artifact of the technique used for the estimates [*Tsimplis et al.*, 2006]. Omitting the oldest estimate of *Schott* [1915], for instance, the inflow range reduces to between 0.72 and 1.26 Sv.

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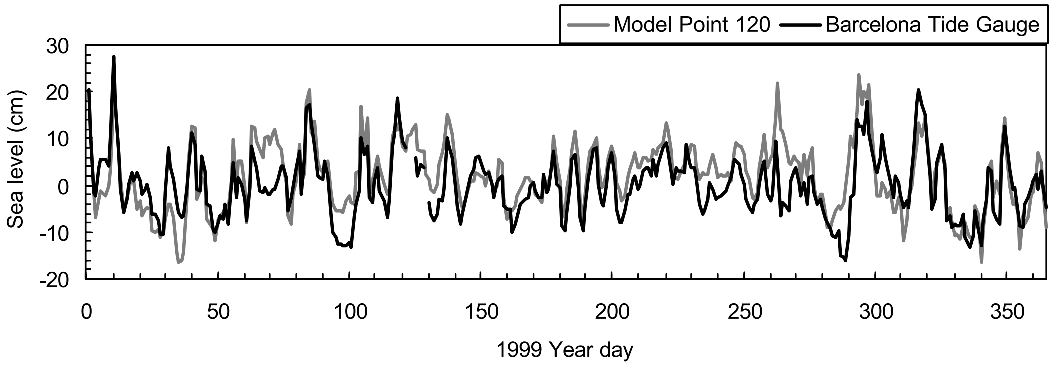


**Figure 1.** Sea level residuals from the Barcelona tide gauge (41°21′01″N, 2°9′41″E) and model values at the closest grid point (41°20′00″N, 2°15′00″E) during 1999.

[4] The referred uncertainties are due to several reasons. First, most referred estimates are based on short (less than 1 year) deployments of moorings, the only exception being the 2 years long record obtained by *Candela* [2001]. Second, the estimates often involve several assumptions about the current structure both in the vertical and across the strait [e.g., *Bryden et al.*, 1994]. These assumptions are required because the accurate estimation of the outflow depends on knowledge of the level of the interface layer [*Tsimplis and Bryden*, 2000]. Last, models have only recently been successful in simulating the three dimensional character of the strait [*Sannino et al.*, 2002].

[5] Since a large part of the variability of the exchanged flow through the Strait of Gibraltar is barotropic in character [*Candela et al.*, 1989], it is worth testing the ability of a two dimensional model driven by atmospheric forcing alone to simulate the variability of the net exchange at subinertial frequencies. We estimate the flow across the Strait of Gibraltar from the available model output as the time derivative of sea level integrated over the whole Mediterranean basin (section 2). The derived flow will be first compared with previously published estimates [*Tsimplis and Bryden*, 2000; *García-Lafuente et al.*, 2002a] focusing on intraseasonal scales (section 3.1). Next, we will focus on the annual cycle (section 3.2), comparing the contribution of the direct atmospheric forcing with the evaporation-precipitation (E-P) budget. At longer timescales, the flow forced by atmospheric pressure and wind is expected to be much smaller than the E-P induced flow.

[6] We further explore the relationship between the modeled sea level and transports, on one hand and regional climatic indices on the other hand, in particular the North Atlantic Oscillation (NAO) and the Mediterranean Oscillation index (MOI) (section 4.1). On the basis of the established relationships we reconstruct the Mediterranean sea level variability for the period 1659–2001 at a monthly basis (section 4.2). We use the reconstructed sea level time series to assess the impacts that the direct meteorological forcing has on the estimates of sea level trends for Mediterranean records with variable lengths.

## 2. Data and Methodology

### 2.1. Model Sea Level Data

[7] Mediterranean sea level has been modeled for the period 1958–2001 in the framework of the HIPOCAS (Hindcast of Dynamic Processes of the Ocean and Coastal Areas of Europe) project [*Guedes Soares et al.*, 2002]. The model is a barotropic version of the Hamburg Shelf Circulation Model (HAMSOM) forced only by atmospheric pressure and wind, and is routinely operated by Puertos del Estado as part of the Spanish sea level forecasting system [*Álvarez-Fanjul et al.*, 1997, 2001]. The covered domain is from 30°N to 47°N and from 12°W to 35°E, with a spatial resolution of 1/4° × 1/6°. In time, the hourly outputs of the model were averaged into daily values for the purpose of this work. The atmospheric pressure and wind fields were produced through dynamical downscaling from the NCEP/NCAR global reanalysis using the atmospheric limited area model REMO [*Sotillo et al.*, 2005]. The resulting winds agree much better with ground observations than the original NCEP/NCAR data, while the pressure field does not present major changes with respect to previous reanalysis.

[8] Regarding the spatial resolution of the model, it is worth stating that (1) it is clearly sufficient to deal with the typical spatial scales of atmospherically forced variability, in particular at timescales longer than a week. EOF analyses have repeatedly shown the high coherency in the basin at these timescales, and (2) regarding sea level adjustment under varying atmospheric forcing, the main role of the strait is to act as a bottleneck. From this point of view, a unique point reproducing the strait is enough, as far as the cross area (the crucial topographic property) associated to the grid point is realistic. The depth of the model grid point representing the Strait of Gibraltar was set to 160 m in order to ensure that the cross section of the strait was accurately represented [*García-Lafuente et al.*, 2002b]. The very good agreement between predicted and observed sea level at different test harbors of the basin confirmed the proper dimensioning of the cross area. The described set up corresponds to the operational version of the model and not to any particular set up for this work. That is, the comparison between model results and transport observations shown later on must be considered as a truly independent test of the model performance.

[9] Prior to this work, the ability of the model to reproduce actual sea level had been largely validated against tide gauge data. Correlation between tide gauge sea level residuals (which include the steric part) and model outputs is typically around 0.8 or higher, and RMS errors are of order 5 cm. As an example (Figure 1) we have plotted the residual sea level (corrected for the tidal signal) recorded by the Barcelona tide

gauge against sea level values hindcasted by the model at the closest grid point (model point 120, located at about 7.5 km from the tide gauge). The agreement between the modeled and actual sea level is very good, and removing the steric part from the observed residuals would be even better. The skill of the model for the complete set of hindcasted parameters is fully evaluated by A.W. Ratsimandresy et al. (A 44-year high-resolution ocean and atmospheric hindcast for the Mediterranean basin developed within the HIPOCAS Project, submitted to *Coastal Engineering*, 2006).

### 2.2. Computation of the Flow Through the Strait of Gibraltar From Mean Sea Level Series

[10] Model sea level values can be integrated over the Mediterranean basin (taking into account the area variation due to latitude) to yield a water volume time series for the basin. Total volume changes, $dV(t)/dt$, would obey an equation like:

$$dV(t)/dt = T(t) + [P(t) - E(t)] + dV_{st}(t)/dt \tag{1}$$

$T(t)$ reflects the change of mass in the Mediterranean basin driven by the net flow through connections with the open ocean. This term is largely dominated by the flow through the Strait of Gibraltar, which will be the only one considered in this work. The term $[P(t) - E(t)]$ is also associated with a mass exchange, but in this case with the atmosphere (Precipitation minus Evaporation, river runoff being included in $P$). The last term on the right-hand side represents steric variations (changes in volume but not in mass), mainly associated with the seasonal heat flux cycle.

[11] Since the model is only driven by the atmospheric mechanical forcing (i.e., it does neither contain the steric component, nor the E-P cycle), then the volume changes computed from the model $dV_m(t)/dt$ must be exclusively linked to the exchanges through the strait. However, this does not mean that the model can recover the total variability of $T(t)$. At timescales between a day and several weeks, the flow through the strait is mainly driven by the atmospheric forcing [*Candela et al.*, 1989; *Tsimplis and Bryden*, 2000; *García-Lafuente et al.*, 2002b], indicating that the primary balance for the flow can be approximated by

$$T(t) = dV_m(t)/dt = S\ dz(t)/dt \tag{2}$$

where $S$ is the total area of the basin and $z(t)$ is the basin mean sea level height as given by the model. However, at annual scales, for instance, the flow induced by the E-P variability is of the same order of the atmospherically induced flow. Moreover, the baroclinic flow induced by changes in the density structure of the water column around the strait is neither included in (2). These limitations must be kept in mind when dealing with long timescales.

[12] Because later in this work we will be interested in estimating the flow from monthly and annual mean sea level, we will first address the accuracy of such estimates. According to (2) the mean flow ($Tp$) over a certain time period $P$ will be given by

$$Tp = P^{-1}\int_P T(t)dt = SP^{-1}\int_P dz(t) = SP^{-1}[z(t+P) - z(t)] \tag{3}$$

This is, the mean flow should be computed simply as the difference between sea level values corresponding to the beginning and the end of the averaging period. For the subinertial variability, mean daily values are used. However, the reconstruction of sea level records from climatic indices cannot be undertaken on a daily basis, but on a monthly basis at most. Therefore a key question is the extent to which the monthly mean flow $Tm$ can be approached by the time derivative of monthly mean sea level $\langle z\rangle_m$:

$$Tm^* = SM^{-1}\left[\langle z\rangle_{m+1} - \langle z\rangle_m\right] \tag{4}$$

where $M$ is a one month time lag. The difference between using (3) and (4) is given by

$$RMS^2 = \left\langle\left\langle (Tm - Tm^*)^2 \right\rangle\right\rangle$$

where $\ll\ \gg$ denote a statistically significant average. Developing this expression and reorganizing terms yields:

$$RMS^2 = S^2M^{-2}\langle\langle\{[z(t+1M) - \langle z\rangle_{m+1}] - [z(t) - \langle z\rangle_m]\}^2\rangle\rangle$$

where $z(t) - \langle z\rangle_m = z'(t)$ is the difference between an instantaneous value and a monthly mean (with the monthly average centered at the time of the instantaneous value). Therefore

$$\begin{aligned} RMS^2 &= S^2M^{-2}\left\langle\left\langle z'(t+1M)^2 + z'(t)^2 - 2z'(t+1M)z'(t)\right\rangle\right\rangle \\ &= 2S^2M^{-2}\left[\sigma_{zm}^2 - \nu_{zm}(1M)\right] \end{aligned}$$

where $\sigma_{zm}^2$ is the variance of the departure of sea level values with respect to their monthly means and $\nu_{zm}(1M)$ is the lag covariance of these departures at 1 month time lag. The expression above can also be written in terms of the lag correlation $C_{zm}$:

$$RMS^2 = 2S^2\ M^{-2}\sigma_{zm}^2[1 - C_{zm}(1M)] \tag{5}$$

Finally, it is useful to express the deviations or errors of our estimate relative to the variability of the series we pretend to reproduce. The latter is given by (3), so that we can compute its variance:

$$\sigma_{Tm}^2 = S^2M^{-2}\left\langle\left\langle [z(t+1M) - z(t) - \langle\langle z(t+1M) - z(t)\rangle\rangle]^2\right\rangle\right\rangle$$

The term $\left\langle\left\langle z(t+1M) - z(t)\right\rangle\right\rangle$ will not be strictly zero if there is a long-term trend in z(t). However, the value of that term is the sea level rise or drop during a month associated with the long trend, which is negligible in front of the other terms. Therefore

$$\sigma_{Tm}^2 \approx S^2M^{-2}\left\langle\left\langle [z(t+1M) - z(t)]^2\right\rangle\right\rangle = 2S^2M^{-2}\sigma_z^2[1 - C_z(1M)] \tag{6}$$

where $\sigma_z^2$ is now the actual variance (not the variance of the departure of sea level values with respect to their monthly means, as it was $\sigma_{zm}^2$) and $C_z(1M)$ is the actual lag

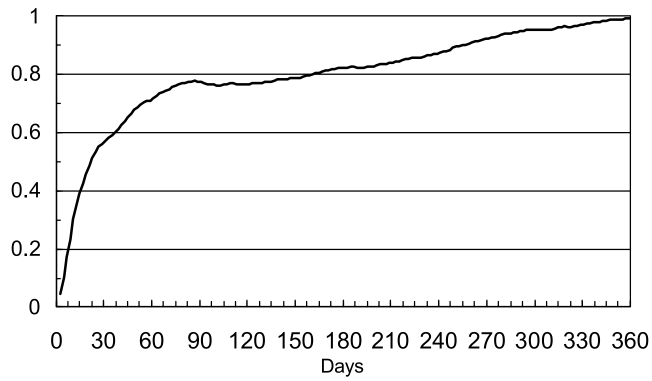


**Figure 2.** Error-to-signal fraction (i.e., error variance divided by the variance of the series to be recovered) associated with the computation of transports from mean sea level values. The plot covers the range from 1 day means to 1 year means.

correlation at 1 month time lag. Therefore a measure of the deviation or error of our estimate expressed as a "noise-to-signal" ratio is given by

$$\eta \equiv RMS^2/\sigma_{Tm}^2 = \left\{\sigma_{zm}^2[1 - C_{zm}(1M)]\right\}\left\{\sigma_z^2[1 - C_z(1M)]\right\}^{-1} \quad (7)$$

Expression (7) is valid for any averaging period and not only for monthly means. We used the 44 years of model sea level data to obtain $\sigma_{zp}^2$ and $C_{zp}(1P)$ for $P$ ranging between 1 day and 1 year. The resulting values of $\eta$ are shown in Figure 2. For short averaging periods $\sigma_{zp}^2 \ll \sigma_z^2$, so that $\eta$ approaches zero. Instead, for long averaging periods $\sigma_{zp}^2 \approx \sigma_z^2$, but since $C_{zp}(P) \approx C_z(P) \ll 1$ ($\approx$0.1 for $P = 1$ yr, for instance), then $\eta \approx 1$. This result suggests that it will make little sense to compute the flow from annual sea level means. For monthly means we will have a significant degree of uncertainty ($\eta = 0.57$), but it will not prevent us from having an acceptable correlation between the estimated and actual flow.

[13] As an example, we show a comparison between *Tm* and *Tm** for both, monthly and annual means (Figure 3). Figure 3 (top) reveals that the flow computed from monthly mean sea level *Tm** follows the actual monthly variability *Tm* quite well, the main deviations being due to an underestimation of the oscillations. Instead, the flow computed from annual mean sea level *Ty** not only underestimates the actual annual mean flow *Ty*, but also the variability between actual and estimated flow is rather different. These results emphasize the convenience of estimating mean transport values from sea level monthly means at most.

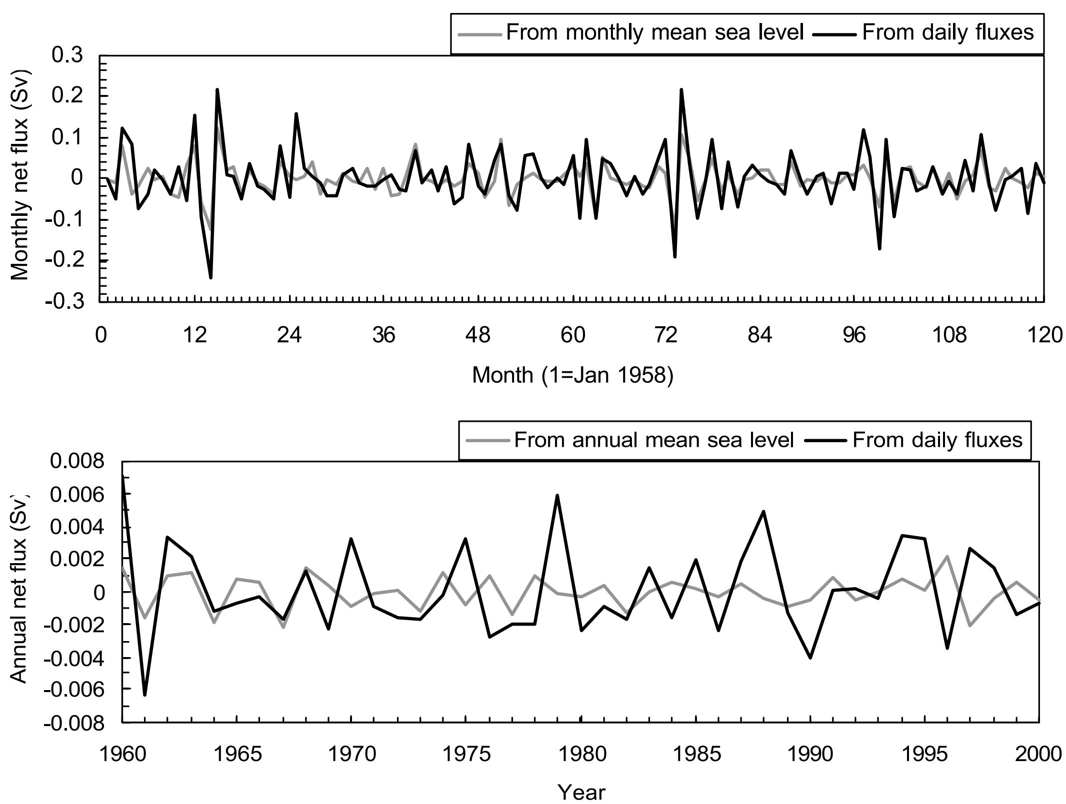


**Figure 3.** Comparison between the mean flow computed as the average of daily flow values and that computed from mean sea level values. (top) Monthly means (only the first 10 years of the modeled period have been plotted). (bottom) Annual means.

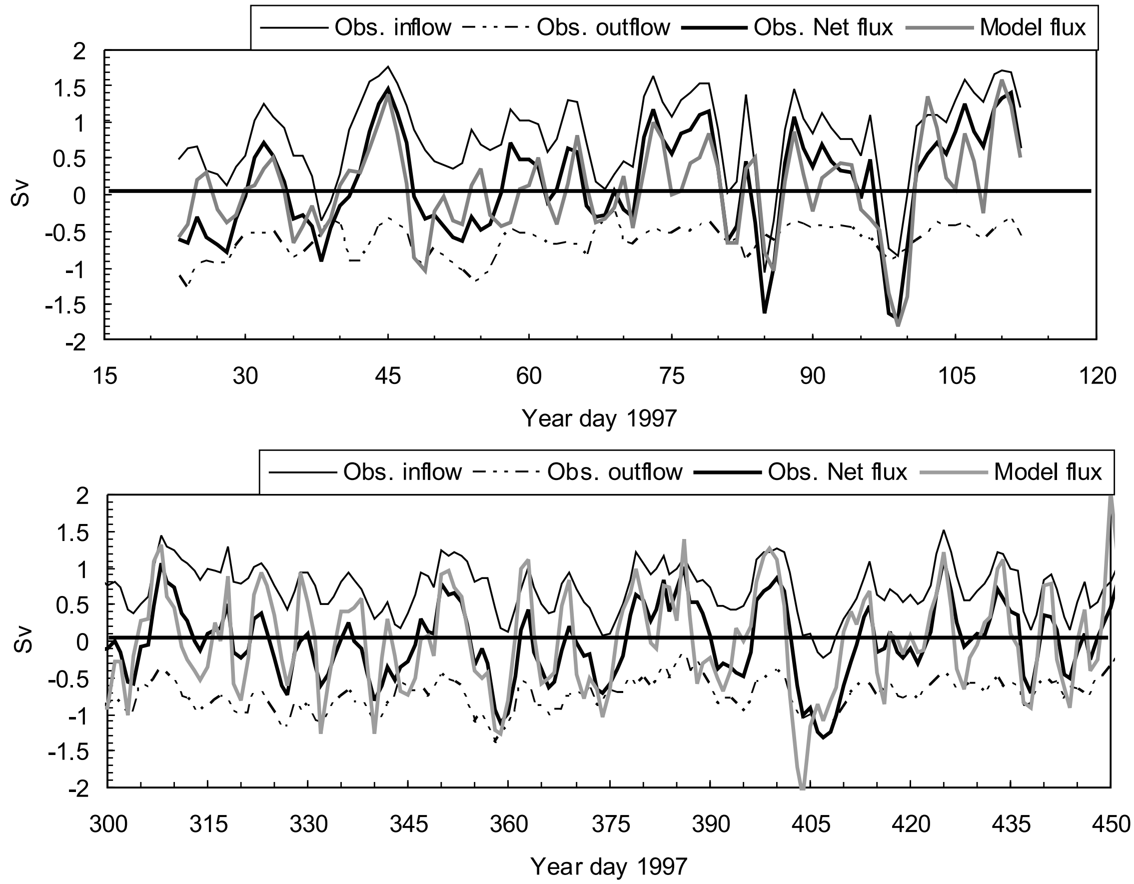


**Figure 4.** Comparison between model-derived flows and those observed by (top) *Tsimplis and Bryden* [2000] (23 January to 23 April 1997) and (bottom) *García-Lafuente et al.* [2002a] (26 October 1997 to 27 March 1998). The observed inflow and outflow have also been plotted.

### 2.3. Observed Flows

[14] Daily flow estimated from the model will be compared with two previously published independent data sets. The first one consists of transport estimates of the inflow and outflow though the Strait of Gibraltar as calculated by *Tsimplis and Bryden* [2000]. They are based on one upward looking ADCP and two moorings located at the sill section of the strait and span from 23 January to 23 April 1997. The second data set consists of the inflow, outflow and net transport values computed by *García-Lafuente et al.* [2002a] for the period from 26 October 1997 to 27 March 1998. In this case the daily transports were estimated from the data provided by a current meter array deployed at the eastern part of the strait.

### 2.4. Climatic Indices

[15] The North Atlantic Oscillation Index used for the model period (1958–2001) was obtained from the Climate Research Unit (http://www.cru.ac.uk) [*Jones et al.*, 1997]. To extrapolate sea level backward in time we used the NAO reconstructions of *Luterbacher et al.* [1999, 2002], which extend back to 1659. Finally, a Mediterranean Oscillation Index (MOI) spanning the modeled period was kindly provided by Kay Šušelj. The MOI is defined as the pressure differences from an area larger than the Mediterranean (from mid-North Atlantic to southeast Mediterranean) and it has been reported to be more closely related to Mediterranean variability than the NAO index [*Supic et al.*, 2004].

## 3. Flow Through the Strait of Gibraltar

### 3.1. Intraseasonal Variability

[16] In Figure 4 the daily flow derived from model sea level is compared with the published transport estimates of *Tsimplis and Bryden* [2000] and *García-Lafuente et al.* [2002a]. For the first case (Figure 4, top) the agreement between model and observation-based values is very good, the correlation between the two series being $r = 0.81$. A linear regression between the two series ($Tobs = a*Tmod + c$) gives a coefficient $a = 0.91$. The mean observed inflow is 0.789 Sv and the mean observed outflow is −0.634 Sv, so that the mean observed net flow is 0.156 Sv (with a standard deviation of 0.696 Sv). The mean net flow derived from model sea level is 0.071 Sv, so that there is a bias of 0.085 Sv between the two series which could correspond to the contributions neglected in (2).

[17] For the second case (Figure 4, bottom), the correlation is still high ($r = 0.77$) but the regression coefficient is poorer ($a = 0.57$). The modeled and observed flows have a very similar variability, however the model overestimates the flow computed by *García-Lafuente et al.* [2002a]. For this period the mean observed inflow is 0.724 Sv and the mean observed outflow is −0.741 Sv, so that the mean

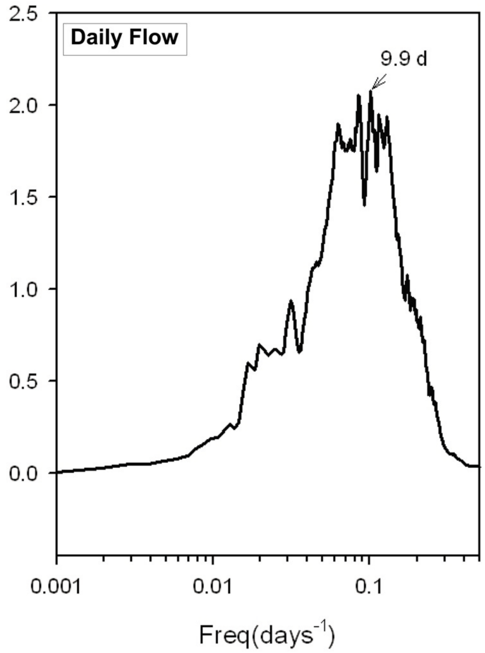


**Figure 5.** Power spectrum of the daily transport through the Strait of Gibraltar estimated by the model for the period 1958–2001.

observed net flow is negative: −0.018 Sv (with a standard deviation of 0.502 Sv). The mean model net flow is also slightly negative (−0.004 Sv), so that the bias between the two series is of −0.014 Sv.

[18] The good agreement between the observed and modeled flows suggests that model daily values can be considered a good approximation to the intraseasonal variability of the actual net flow. Hence we can infer the basic characteristics of sea level and the resulting net exchange for the whole period covered by HIPOCAS (1958–2001). During that period, the response of the basin mean sea level to atmospheric pressure and wind forcing ranges between ±20 cm with a standard deviation of 5.5 cm. The derived flow ranges between ±2.7 Sv except for an isolated value of 3.8 Sv, and has a standard deviation of 0.56 Sv.

[19] A spectral analysis of the derived flow shows a dominant broad peak covering frequencies between one and two weeks, with a sudden drop for larger and smaller frequencies (Figure 5). In particular, we note that we do not observe a 5.5 day oscillation corresponding to the Helmholtz mode of the Mediterranean Sea, as computed by the model of *Candela et al.* [1989]. Harmonic signals at 15 and 30 days are detectable, but they are rather small (0.0025 and 0.004 Sv respectively). The 15 and 30 day oscillations are not a result of the ordinary astronomical tides although one cannot exclude the possibility of being partly linked with the atmospheric tidal contribution at these frequencies. The direct astronomical forcing on the ocean is not included in the model, which could explain the discrepancy with the results of *Tsimplis and Bryden* [2000], who suggest values two orders of magnitude higher for the inflow and the outflow at these frequencies. In any case, the good agreement between the model and the observation-based flows (Figure 4) indicates that the net effect of these harmonics is not significant for the net flow.

### 3.2. Annual Cycle

[20] The annual sea level and transport cycles have been estimated by averaging the values corresponding to the same year day over the 44 years (Figure 6, top). The contribution of the meteorological forcing to the total sea level cycle has a maximum value in April and a minimum value in October, the range of the annual cycle being about ±3 cm. The associated transport values are mostly in the range of ±0.2 Sv, but it is not easy to distinguish an annual cycle from the daily series of Figure 6 (top). Monthly means computed from the daily values of Figure 6 (top) show a transport cycle consistent with the annual sea level cycle: it has a maximum value in March and a minimum value in September (Figure 6, bottom). The range of the annual cycle in the transport is ±0.025 Sv, with a mean value of less than 0.005 Sv. The range of the annual sea level cycle computed on the basis of monthly means is ±2.3 cm, which is slightly smaller than that calculated on the basis of daily values.

[21] The annual cycle of the volume change induced by the evaporation-precipitation budget is also plotted in Figure 6 (bottom). This has been computed using monthly E-P values from the NCEP/NCAR reanalysis for the period 1949–2001 (see *Josey* [2003] for further discussion of the NCEP/NCAR E-P fields in the Mediterranean Sea). The volume change induced by the E-P budget is always positive and has a mean value of 0.065 Sv, corresponding to a mean fresh water deficit of 75 cm year$^{-1}$. This estimate is larger than the 0.04 Sv reported by *Garrett et al.* [1993]. A similar discrepancy with the E-P budget of *Garrett et al.* [1993] is reported by *Candela* [2001].

[22] The E-P contribution to the annual sea level cycle is about three times larger than the contribution of atmospheric pressure and wind However, the amplitude of the two cycles with respect to their respective annual means is very similar (±0.025 Sv). Hence, assuming that the E-P cycle results in a net flow through the strait that exactly compensates for the water deficit (at least on an annual basis), the annual cycle of the total flow will be significantly modulated by the direct meteorological forcing. The total annual cycle will show for instance a late winter relative maximum of 0.072 Sv, not much smaller than the absolute maximum reached in late summer (0.080 Sv). *García-Lafuente et al.* [2002c] report a semiannual signal in the net flow through Gibraltar of 0.064 ± 0.035 Sv with a phase of 187 ± 0.035 degrees (peaking late March and late September), in agreement with these findings. The associated sea level spring peak has also been observed in other tide gauge records and it has been associated with a semiannual cycle of total sea level [*Bryden et al.*, 1994; *García-Lafuente et al.*, 2004; *Tsimplis et al.*, 2005].

[23] *Bryden et al.* [1994] suggest an annual cycle for the inflow of 0.12 Sv with the maximum on day 261. For the outflow they report an annual cycle of 0.03 Sv with maximum on day 216, which added to the inflow gives almost 0.15 Sv around year day 257 for the net flow. Although the location of the maximum is compatible with the broad maximum of the E-P budget shown in Figure 6, the flow values given by *Bryden et al.* [1994] seem unrealistically high. On the other hand, *García-Lafuente et*

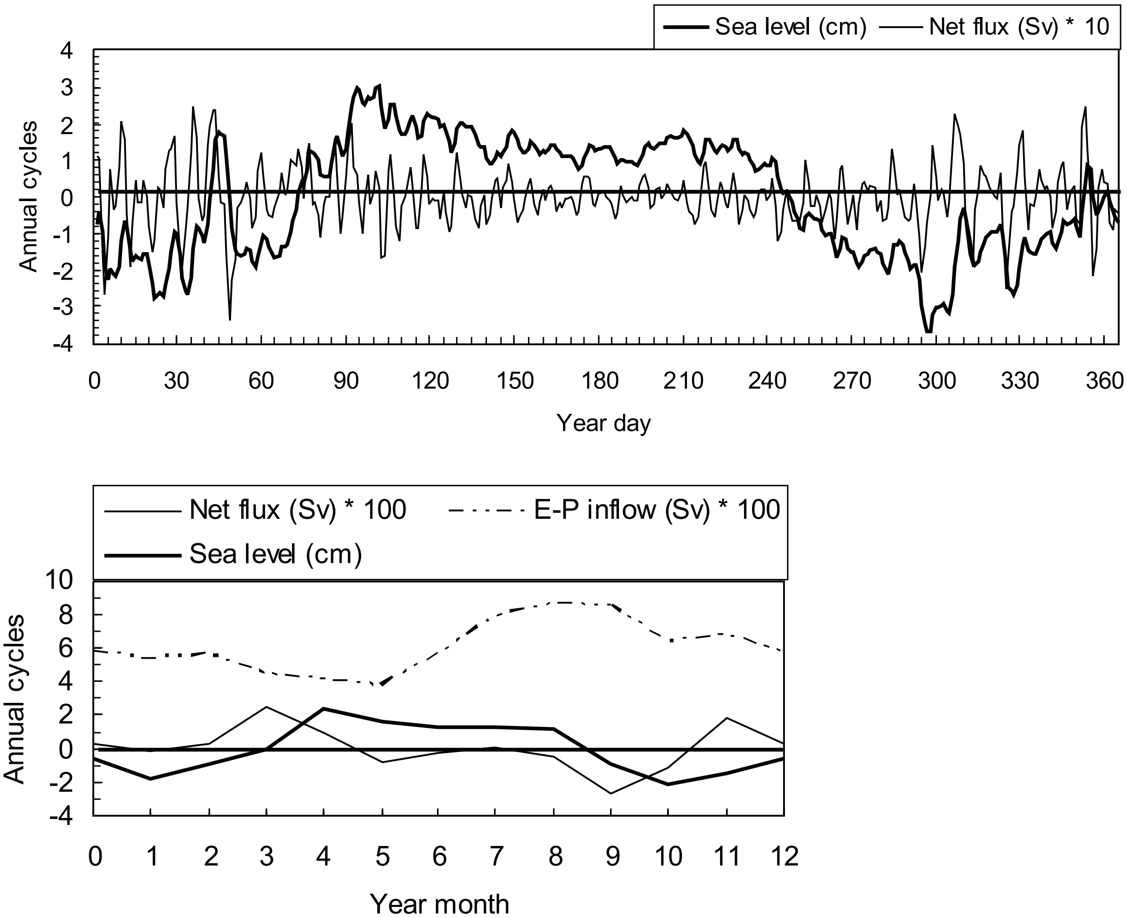


**Figure 6.** Annual cycles of model sea level and the derived flow through the Strait of Gibraltar. (top) Computed on a daily basis. (bottom) Monthly means of the above daily values, with the Mediterranean volume change inferred from the monthly E-P budget overplotted.

*al.* [2002c] computed an annual cycle of 0.077 ± 0.044 Sv, with a maximum on year day 237 ± 33 for the net flow. These values are in rather good agreement with the cycles shown in Figure 6, and the location of the maximum is right at the centre of the broad maximum of the E-P budget.

[24] For longer timescales the meteorologically forced flow through the Strait of Gibraltar is very small. As stated above, annual mean flow values are of order $10^{-3}$ Sv. When averaging over the four decades spanned by the model, mean values are of order $10^{-4}$–$10^{-5}$ Sv, which is consistent with the trend of about −0.6 mm/yr observed in the direct meteorological forcing of sea level [*Tsimplis et al.*, 2005].

## 4. Reconstruction of Sea Level and Transport

### 4.1. Correlation Between Sea Level and Climatic Indices

[25] The model flow is obtained as the first time derivative of total sea level volume in the basin. Therefore, if a good proxy of the Mediterranean sea level variability can be used to reconstruct the mean sea level, we could then reproduce the flow through the strait through its derivative. In section 2 we have shown that a reliable flow can be estimated at most from monthly means, but not from annual means. Therefore the regression between sea level and climatic indices will be undertaken on a monthly basis.

[26] The NAO and MOI indices are both known to correlate well with Mediterranean Sea level, either recorded by tide gauges [e.g., *Tsimplis and Josey*, 2001] or inferred from altimetry [*Woolf et al.*, 2003]. The western Mediterranean and the Adriatic appear better correlated to the NAO than the eastern Mediterranean as they are closer to the centers of action of the pressure systems. In Figure 7 the two indices are plotted against the basin-wide monthly mean sea level computed from model data. The correlation coefficient with the NAO is 0.58 with a regression coefficient of 1.14 cm/unit NAO while the MOI gives a correlation of 0.86 with a regression coefficient of 2.92 cm/MOI index. On an annual basis the correlation with NAO is slightly smaller (0.55) whereas with the MOI is slightly larger (0.91). During winter (December–March) the correlation with the NAO increases up to 0.86 and with the MOI is even better.

[27] The better correlation with the MOI and winter NAO index is partly due to the fact that during the modeled period both indices present a trend similar to that observed for sea level (−0.6 mm/yr [*Tsimplis et al.*, 2005]), whereas the NAO index show a smaller trend. Hence a much better fit is obtained when model sea level is bilinearly regressed with climatic indices plus a linear trend ($Z(t) = a*Ind + b*t + c$). In this case, the correlation with the NAO index increases to

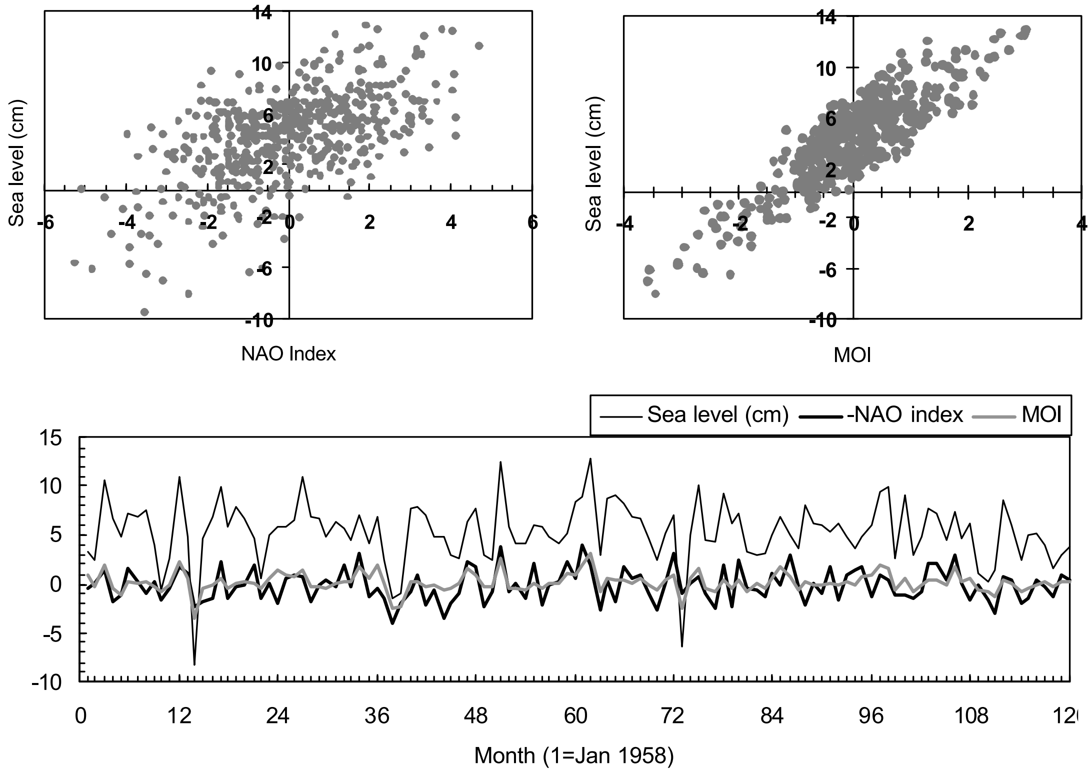


**Figure 7.** (top) Scatterplots between monthly mean sea level and monthly climatic indices (NAO and MOI) for the whole modeled period (1958–2001). (bottom) Sample (first 10 years of the modeled period) of monthly mean sea level and monthly climatic indices.

0.626 for monthly values and up to 0.84 for annual values. The correlation with the MOI does not significantly increase for monthly values (from 0.86 to 0.87) but it does for annual values (from 0.91 to 0.95). However, the extrapolation of the observed trend backward in time presents some conceptual problems. Moreover the impact of such a trend on the monthly and even annual variability of the estimated flow is expected to be very small. Therefore we based the reconstruction on a regression of sea level with climatic indices only.

[28] Despite the better correlation between the MOI and the basin mean sea level we had to use the NAO further on as there are not any reconstructions of the MOI known to us. Thus we used the *Jones et al.* [1997] NAO index to reproduce the Mediterranean Sea level over the period of instrumental observations (from 1821 onward). In addition, we used the NAO reconstruction of *Luterbacher et al.* [1999, 2002] for which we determine a separate regression coefficient (−2.22 cm/unit NAO, $r = 0.74$) on which we reconstructed the Mediterranean Sea level since 1659.

### 4.2. Mediterranean Sea Level Variability and Flow for the Period 1659–2001

[29] In Figure 8 the monthly sea level and transport for the decade 1991–2000 are plotted over the estimates based on the NAO regression. It can be seen that using the NAO index as a proxy allows the recovery of a significant part of the variability of sea level and the associated flow, although both reconstructed series underestimate actual values. In the case of the flow through the Strait of Gibraltar, part of the underestimation can be attributed to the fact that they are computed from monthly mean sea level, as pointed out in section 2 (Figure 3).

[30] Assuming that the modeled period (1958–2001) is representative of the whole reconstructed period, the accuracy of the reconstruction can be assessed by comparing the reconstructed with the modeled values over the 44 years of model data. The RMS difference between reconstructed and actual model values is 2.7 cm for sea level and 0.061 Sv for the flow.

[31] The characterization of the monthly reconstructed series can be summarized as follows: monthly mean sea level ranges between ±5 cm whereas the net flow keeps in between ±0.05 Sv. Both series show a progressive variance decrease as they extend backward in time, in particular before 1750. This is a direct consequence of the variance decrease of the NAO index reconstructed by *Luterbacher et al.* [1999, 2002]. More interesting than the characterization of the monthly series is to investigate the decadal and centennial variability. This will be done for sea level only, as the uncertainty in the computation of the net flow is larger than its variance for such long timescales.

[32] In order to explore the decadal and centennial variability we run a 1 yr and a 10 yr moving average through the reconstructed monthly sea level data extending back to 1659. For these series, the deviation between regressed and actual sea level is 0.41 and 0.22 cm respectively. The 10 year average together with a 10 year average

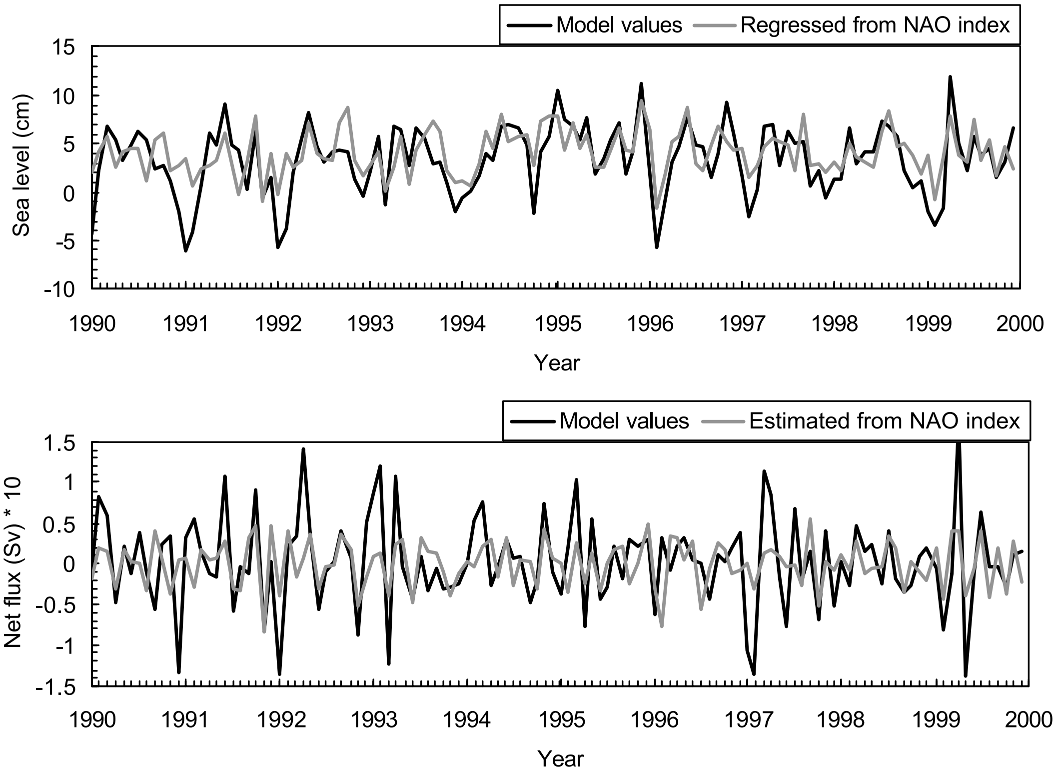


**Figure 8.** (top) Comparison between monthly mean sea level given by the model and the values regressed from the NAO index for the decade 1990–2000. (bottom) As in Figure 8 (top) but for the monthly mean flow through the Strait of Gibraltar.

of the reconstructed winter mean sea level is shown in Figure 9. For the latter, the deviation between the regressed and actual sea level (averaged over the 44 years of model data) is 0.76 cm. The deviations are therefore significantly larger than for the record spanning the four seasons, but since the standard deviation of the series is also much larger, the relative accuracy of the reconstructions is similar for both series. Similar analysis of the flows demonstrated that the uncertainty in the computation of the net flow is larger than its variance for such long timescales. Thus these results are not presented.

[33] The reconstructed sea level records do not include any information on steric changes or water mass additions or extractions but are purely forced by direct atmospheric forcing. However, trends and interdecadal variability is evident in the reconstructed records. Thus the existence of

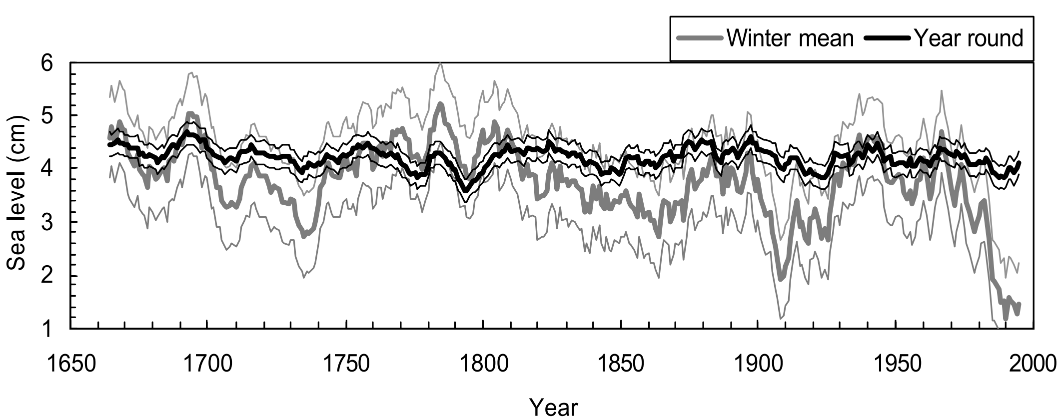


**Figure 9.** Ten-year moving average of the monthly sea level series reconstructed from the NAO index and of winter (December–March) mean sea level reconstructed from the winter NAO index. The uncertainty associated with the reconstruction of both series has been plotted.

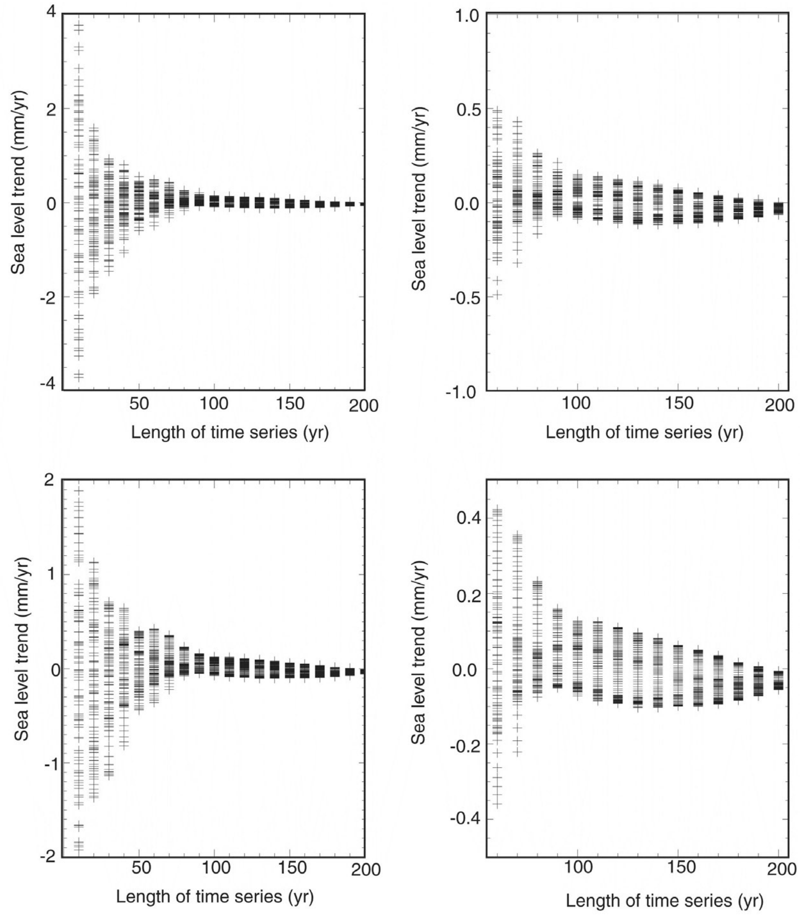


**Figure 10.** Sea level trends calculated for the reconstructed time series by selecting parts of different lengths. (top) Results from the reconstructed winter sea level (left) with the longer than 50 year segments repeated in larger scale (right). (bottom) Results for the 10 year filtered values.

century scale trends in the response of sea level to direct meteorological forcing can be assessed on the basis of the reconstructed sea level. The year-round series shows only a very small trend (−0.005 mm/yr), while the winter mean sea level shows a more apparent trend (−0.033 mm/yr), in particular after 1800. By selecting specific periods one can maximize or minimize the trend estimates. Thus we have calculated the trends for the following periods, according to the major features seen in Figure 9: from the beginning to 1740, from 1740 to 1810, from 1820 to 1900, from 1900 to 1959 and from 1960 to 2000. The corresponding trends are −0.26, 0.06, 0.03, 0.3 and −0.9 mm/yr for winter sea level and −0.08, 0.02, 0.01, 0.13 and −0.31 mm/yr for the whole year. Thus if one was observing sea level during the period 1820−1900 and then again between 1900 and 1960, the result would be a change in sea level rise from 0.03 to 0.3 mm/yr for the winters and about a third of this for the whole year. With standard errors from the regression of about 0.04 and 0.07 mm/yr, a statistically significant change of the trends would have been claimed around 1900.

[34] The previous results can be generalized in the following way: we calculate the trends for segments of the record of 10 to 200 years in multiples of 10 years starting from the first year and then we increase the starting point by one year and repeat the calculation. The resulting distribution of sea level trends for the various segments is shown in Figure 10. The range of trends obtained for the 10 year segments range from −5.56 to 5.86 mm/yr while for the filtered time series the range is −2.6 to 1.9 mm/yr. For 50 yearlong records the range of values obtained are in the range of −0.55 to 0.58 mm/yr and for records 100 yr long the range is −0.18 to 0.15 mm/yr for the unfiltered values. This means that in an extreme situation the values of sea level trends in the Mediterranean Sea between two centuries may differ up to 0.33 mm/yr with these values being part of the change induced by direct meteorological forcing. As a consequence, any other factor, like heat expansion or mass addition due to global warming would need to cause sea level rise exceeding this level of acceleration in order to be detectable.

## 5. Conclusions

[35] We have shown that the subinertial variability of the barotropic flow through the Strait of Gibraltar can be recovered with good accuracy from the Mediterranean mean sea level provided by a barotropic model forced with atmospheric pressure and wind only. This has been demonstrated by a direct comparison between the flow derived from model sea level and that estimated from current meter

observations for two previously published independent data sets. It is worth noting that the measurements correspond not only to different periods but also to different locations at the strait (the first were obtained at the Camarinal Sill section and the second at the eastern part of the strait, within the Mediterranean). They also involved different interpolation techniques. The good agreement between the modeled and observed transports confirms that for intraseasonal scales, the meteorological forcing dominates the variability of the flow through the strait, in particular for periods between one and two weeks. On this basis, we have been able to characterize the net flow for the period 1958–2001, giving maximum ranges as well as its standard deviation.

[36] At annual timescales, the mean value of the model derived net flow through the Strait of Gibraltar is about 0.005 Sv, with a standard deviation of ±0.025 Sv. This variability is similar to the variability of the volume change derived from the negative evaporation-precipitation budget, though the latter has a mean value of about 0.065 Sv. Hence the atmospheric flow significantly modulates the annual cycle of the incoming Atlantic water, producing a late winter peak associated with a sea level secondary maximum observed both in tide gauge records and in previously computed flows [*Bryden et al.*, 1994; *García-Lafuente et al.*, 2002c; *Tsimplis et al.*, 2005]. Thus we confirm earlier suggestions made by *Tsimplis and Woodworth* [1994] and *García-Lafuente et al.* [2004] that the semiannual cycle observed in sea level records is due to the direct atmospheric forcing. In addition, we have been able to separate and quantify the contribution of the mechanical atmospheric forcing from the annual cycle induced by the E-P budget. At inter annual and longer than annual timescales the meteorological contribution accounts for a small fraction of the variability of the net flow through the Strait of Gibraltar.

[37] We have reconstructed monthly sea level and barotropic flow time series for the period 1659–1957 based on statistically significant correlations between the NAO and the Mediterranean Sea level. However, the barotropic flow time series were found to have significantly larger uncertainty than the daily series of the period 1958–2001, thus precluding any meaningful results from the reconstructed transports.

[38] By contrast, the reconstructed sea level record has allowed us to examine decadal to centennial scales since 1659. Centennial and decadal variability has been found in the reconstructed sea level time series, which gives rise to trends if the time series is fragmented in appropriate periods. Thus it has been possible to assess that records of up to 50 years long will have a bias of up to 0.6 mm/yr due to the direct atmospheric forcing. As an example, if one was observing sea level during the period 1820–1900 and then between 1900 and 1960, the result would be a change in sea level rise from 0.03 to 0.3 mm/yr for the winters and about a third of this for the whole year. This coincides with the observed acceleration of sea level around the beginning of the century claimed by *Church et al.* [2001]. Taking into account that the NAO is probably more important for northern Europe that for the Mediterranean Sea (especially in the shallow North Sea where wind forcing strongly affects sea level changes [*Wakelin et al.*, 2003]) and that sea level is anticorrelated between northern and southern Europe [*Woolf et al.*, 2003], it is possible that the acceleration observed by *Church et al.* [2001] is higher at least for the stations located at the east coast of the North Sea while for the Atlantic coasts the effects should be of the same sign as in the Mediterranean. However, the above suggestion could be confirmed or rejected by doing similar work in the northern Europe.

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