Atlantic to Mediterranean Sea Level Difference Driven by Winds near Gibraltar Strait

DIMITRIS MENEMENLIS, ICHIRO FUKUMORI, AND TONG LEE

Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California

ABSTRACT

Observations and numerical simulations show that winds near Gibraltar Strait cause an Atlantic Ocean to Mediterranean Sea sea level difference of 20 cm peak to peak with a 3-cm standard deviation for periods of days to years. Theoretical arguments and numerical experiments establish that this wind-driven sea level difference is caused in part by storm surges due to alongshore winds near the North African coastline on the Atlantic side of Gibraltar. The fraction of the Moroccan coastal current offshore of the 284-m isobath is deflected across Gibraltar Strait, west of Camarinal Sill, resulting in a geostrophic surface pressure gradient that contributes to a sea level difference at the stationary limit. The sea level difference is also caused in part by the along-strait wind setup, with a contribution proportional to the along-strait wind stress and to the length of Gibraltar Strait and adjoining regions and inversely proportional to its depth. In the 20–360-day band, average transfer coefficients between the Atlantic–Alboran sea level difference and surface wind stress at 36°N, 6.5°W, estimated from barometrically corrected Ocean Topography Experiment (TOPEX)/Poseidon data and NCEP–NCAR reanalysis data, are 0.10 ± 0.04 m Pa^{-1} with 1 ± 5-day lag and 0.19 ± 0.08 m Pa^{-1} with 5 ± 4-day lag for the zonal and meridional wind stresses, respectively. This transfer function is consistent with equivalent estimates derived from a 1992–2003 high-resolution barotropic simulation forced by the NCEP–NCAR wind stress. The barotropic simulation explains 29% of the observed Atlantic–Alboran sea level difference in the 20–360-day band. In turn, the Alboran and Mediterranean mean sea level time series are highly correlated, $\rho = 0.7$ in the observations and $\rho = 0.8$ in the barotropic simulation, hence providing a pathway for winds near Gibraltar Strait to affect the mean sea level of the entire Mediterranean.

1. Introduction

The sea level difference between the Atlantic Ocean and the Mediterranean Sea has been attributed mostly to the following: tides (e.g., Brandt et al. 2004), atmospheric pressure fluctuations (e.g., Tsimplis and Josey 2001), steric contributions (e.g., Cazenave et al. 2002), and geostrophic or hydraulic controls within the Strait of Gibraltar (e.g., Ross and Garrett 2000). Some studies (Fukumori et al. 2007; García-Lafuente et al. 2002a,b; Garrett 1983), however, suggest that winds in the vicinity of Gibraltar Strait may be a fifth major cause of Atlantic to Mediterranean sea level difference. García-Lafuente et al. (2002a) describe a meteorological forcing event that interrupts inflow in the Strait of Gibraltar. They estimate an approximate contribution from

wind forcing of 0.3 Sv (Sv $\equiv 10^6$ m^3 s^{-1}), which is added to a 0.5–0.6-Sv contribution from atmospheric pressure. Fukumori et al. (2007) analyze altimetric sea level observations, from which tidal and atmospheric pressure signals have been removed, and identify a near-uniform basin-wide sea level fluctuation of the Mediterranean Sea with periods ranging from days to years and with amplitudes of up to 0.2 m peak to peak. They estimate that these fluctuations are caused by a net mass flux through the Strait of Gibraltar with a peak 10-day-averaged amplitude of 0.26 Sv, corresponding to a depth-averaged current of 0.08 m s^{-1} at the Camarinal Sill. By comparison, Astraldi et al. (1999) report that tidal currents within the strait have velocities higher than 2.5 m s^{-1}, that atmospheric pressure fluctuations cause currents with magnitudes of about 0.4 m s^{-1}, and that there are baroclinic currents with magnitudes of about 0.5 m s^{-1} that are induced by the specific budget of the Mediterranean Sea.

From theoretical considerations, Garrett (1983) estimates that wind-driven storm surges on the continental

Corresponding author address: Dimitris Menemenlis, Jet Propulsion Laboratory, California Institute of Technology, Mail Stop 300–323, 4800 Oak Grove Dr., Pasadena, CA 91109.
E-mail: menemenlis@jpl.nasa.gov

shelf of North Africa, just outside the Strait of Gibraltar, may produce sea level changes in the Mediterranean that are up to 30% of the pressure-driven variability. Using numerical simulations, forced first by atmospheric pressure alone and then by atmospheric pressure and wind stress, García-Lafuente et al. (2002b) infer that, although the main driving force of subinertial flow through Gibraltar Strait is atmospheric pressure over the Mediterranean Sea, wind stress on the Atlantic side of the strait may also contribute appreciably to the subinertial net flow. Using numerical experiments in a quasi-global, coarse-resolution ocean circulation model, Fukumori et al. (2007) proposed that the observed basin-wide intraannual sea level variability of the Mediterranean Sea is barotropic in the sense that the fluctuations are insensitive to stratification. They also suggest that the fluctuations are driven by winds in the vicinity of Gibraltar Strait. In particular, their coarse-resolution simulations show the existence of a dynamic balance between the Atlantic–Mediterranean sea level difference and the winds at the Strait of Gibraltar and its neighboring regions, including the Alboran Sea and a part of the Atlantic Ocean immediately to the west of the strait. The present study aims to identify and to describe in detail the physical processes responsible for this dynamic balance.

Understanding and quantifying externally forced sources of spatially correlated sea surface height variability are important to scientists and to engineers that use altimeter data to study other oceanographic processes in the Mediterranean Sea. Fieguth et al. (1998) used a multiresolution optimal interpolation algorithm to estimate the error budget of Mediterranean altimeter data. For barometrically corrected Ocean Topography Experiment (TOPEX)/Poseidon data, they estimate low-wavenumber errors to have a standard deviation of 2.5 cm during the summer months and of 4–8 cm during the winter. If it were possible to explain and to predict some substantial fraction of these errors, more accurate sea level maps could be generated for studies of the Mediterranean dynamics, climate, and sea level variability.

Along-strait winds above Gibraltar are locally intensified (Dorman et al. 1995) and they impact the surface inflow of Atlantic water. Garrett et al. (1989) estimate that the Atlantic inflow current variability caused by along-strait winds is of order 0.8–1.4 m s^{-1} root-mean-square (rms) at the surface. Stanichny et al. (2005) note a high correlation between the zonal wind component and the Tarifa to Ceuta cross-strait sea level difference. They estimate that a 10 m s^{-1} zonal wind will cause a cross-strait sea level difference of 0.85 m corresponding to a zonal surface geostrophic current of order 0.4 m s^{-1}. The impact of along-strait winds on the along-strait sea level difference, however, is assumed to be small (Garrett et al. 1989).

The hypothesis considered herein is that wind-driven Atlantic to Mediterranean sea level difference does not, to first order, occur within the Strait of Gibraltar. Instead, as suggested by Garrett (1983), it is proposed that the sea level difference is primarily caused by winds blowing over the shallow and wide Atlantic continental shelf near the western entrance of Gibraltar Strait. Garrett (1983) argues that the principal cause of the low-frequency, wind-driven, Atlantic–Mediterranean sea level difference is alongshore wind on the continental shelf of North Africa just outside the Strait of Gibraltar. Garrett [1983, Eq. (7.1)] estimates that the sea level difference is approximately 1 cm for a 1 m s^{-1} wind, that is, for a wind stress of approximately 0.002 Pa.

This study revisits the Garrett (1983) hypothesis and estimate using TOPEX/Poseidon sea level data, surface wind stress from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis, and high-resolution numerical simulations of Gibraltar Strait. While the Garrett (1983) estimate is based on the effects of nearshore currents, this study finds large contributions from currents offshore of the 284-m isobath and from the along-strait wind setup. This study also finds that the sea level difference is approximately proportional to the wind stress rather than the wind speed.

The remainder of this article is organized as follows. Section 2 compares the observations of sea level and winds near Gibraltar Strait. Section 3 lays out a theoretical basis for the wind-driven Atlantic to Mediterranean sea level difference. Section 4 presents results from high-resolution barotropic simulations of wind-driven currents through the Strait of Gibraltar. Discussion and conclusions follow in section 5.

2. Observations of sea level and wind near Gibraltar

This section aims to compare time series of the sea level difference between the Atlantic and Mediterranean ends of Gibraltar Strait with surface wind stress above the Atlantic continental shelf immediately west of the strait. Figure 1 depicts the complex bathymetry of Gibraltar Strait and the adjoining seas. On the Atlantic side of the strait there is a wide and shallow continental shelf. By comparison, the Alboran Sea on the eastern side of the strait has a much steeper coastline. The shallowest section of the Strait of Gibraltar is at the Camarinal Sill, near the western entrance, where the maximum water depth is 284 m.
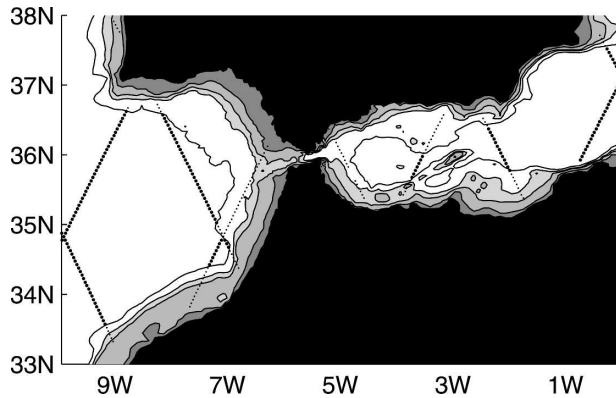
FIG. 1. Bathymetry and TOPEX/Poseidon tracks near Gibraltar Strait. Bathymetric contour lines indicate the 80-, 284-, 500-, and 1000-m depths. The shallowest section of the strait is 284 m deep at the Camarinal Sill. Large dots along the tracks represent altimetric observations, which are used to estimate the sea level on the Atlantic side of Gibraltar Strait and in the Alboran Sea. Small dots represent available but unused altimetric data in regions shallower than 1000 m.

a. TOPEX/Poseidon sea level

Many studies of sea level near and within Gibraltar Strait are based in part on coastal tide gauge data (e.g., Brandt et al. 2004; Cazenave et al. 2002; García-Lafuente et al. 2004; Garrett et al. 1989; Sannino et al. 2004; Stanichny et al. 2005; Tsimplis and Baker 2000). Tide gauge data at Cadiz, Spain, or Tangier, Morocco, are often taken to represent Atlantic sea level while tide gauge data at Malaga, Spain; the British overseas territory of Gibraltar; or the Spanish enclave of Ceuta in North Africa are taken to represent Mediterranean sea level. Coastal sea level stations, however, are not necessarily representative of basin-averaged sea level [e.g., cf. Fig. 3, showing the Cadiz-to-Malaga sea level drop, with Fig. 4b, showing the northeastern Atlantic to western Mediterranean sea level drop, in Ross and Garrett (2000)]. To reduce complications due to near-shore processes, altimetric rather than coastal tide gauge data are used to estimate the Atlantic–Mediterranean sea level difference.

Figure 2 shows time series of the spatially averaged sea level for the Atlantic side of Gibraltar Strait, η_{atl}; for the Alboran Sea, η_{alb}; for the Mediterranean, η_{med}; and for the sea level difference, $h = \eta_{alb} - \eta_{atl}$; see Table 1 for a summary of the notation used. Sea level data are taken from the National Aeronautics and Space Administration (NASA) Goddard Space Flight Center Altimeter Ocean Pathfinder TOPEX/Poseidon sea surface height anomaly, version 9.2, obtained from the NASA Jet Propulsion Laboratory Physical Oceanography Distributed Active Archive Center (informa-

tion available online at http://podaac.jpl.nasa.gov). The Pathfinder altimeter data are derived from TOPEX/Poseidon geophysical data record files, they are interpolated to a specific ground track, and they are provided at 1-s intervals, approximately every 6 km along the ground track. The data cover the period 23 September 1992–11 August 2002, that is, TOPEX/Poseidon cycles 1–364. All known geophysical, media, and instrument corrections have been applied, including corrections for tides and for atmospheric loading.

Fluxes associated with atmospheric loading dominate subinertial variations through Gibraltar Strait (Candela et al. 1989). The altimeter data used herein have been adjusted for atmospheric loading according to the inverse barometer correction (Larnicol et al. 1995; Ponte et al. 1991); that is, sea level drops 1 cm when atmospheric pressure rises by 1 hPa. Deviations from an inverse barometer response in the Mediterranean have been observed, especially for periods shorter than a few days, possibly because of restrictions of flow through Gibraltar Strait (Garrett and Majaess 1984; Le Traon and Gauzelin 1997). For periods of 20 days and longer, however, deviations from the inverse barometer response are expected to be small.

Atlantic and Alboran mean sea level time series are computed based on, respectively, four altimeter tracks west and east of Gibraltar Strait, as shown in Fig. 1. Altimeter data from regions shallower that 1000 m are not included to reduce contamination of the time series by coastal processes and altimeter errors. Mediterranean sea level is computed based on all Mediterranean tracks, including tracks in the Alboran Sea. The black lines in Fig. 2 are constructed based on 364 ten-day averages, 1 ten-day average per TOPEX/Poseidon cycle. Therefore, signals with periods shorter than 20 days cannot be resolved. The gray lines are low-pass-filtered time series with a cutoff period of 360 days. Specifically, the available TOPEX/Poseidon data are first linearly interpolated in time to generate daily time series and then an eighth-order Chebyshev type I filter with a 360-day cutoff period is applied (Antoniou 1993). That is, the gray lines in Fig. 2 represent the annual cycle and interannual variability.

b. NCEP–NCAR surface wind stress

The surface wind stress is from the NCEP–NCAR atmospheric reanalysis (Kistler et al. 2001). Wind stress, rather than 10-m winds, is used in this study because NCEP–NCAR winds are unrealistically low in coastal regions. The quality of the NCEP–NCAR wind stress data near Gibraltar Strait can be assessed through comparison with observations. Dor-
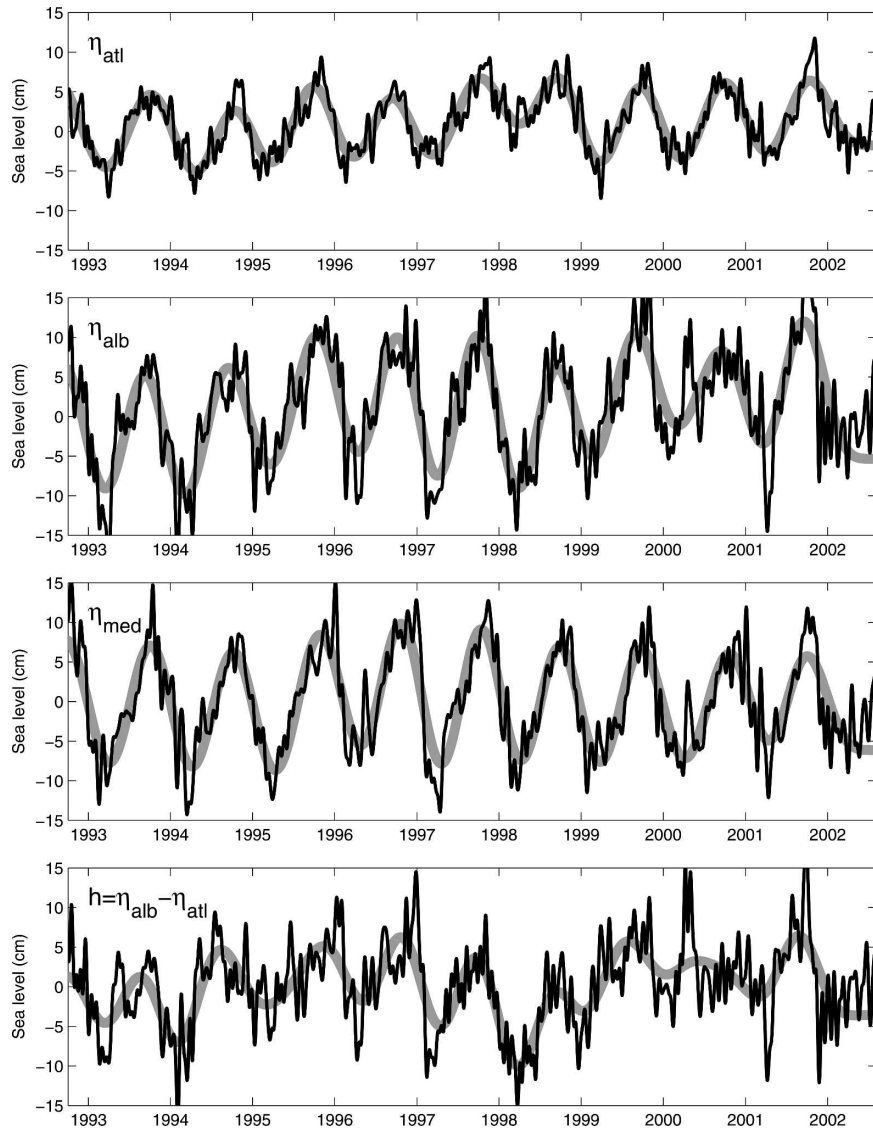
FIG. 2. TOPEX/Poseidon mean sea level time series for the Atlantic side of Gibraltar (η_{atl}), the Alboran Sea (η_{alb}), the Mediterranean (η_{med}), and the Alboran–Atlantic difference ($h = \eta_{alb} - \eta_{atl}$). Black lines represent sea level after application of all standard corrections, including corrections for tides and for atmospheric pressure. Gray lines are low-pass-filtered time series with a cutoff period of 360 days.

man et al. (1995) report along-strait winds above Gibraltar with standard deviations ranging from 7.5 m s^{-1} in Punta Cires, Morocco, to 11.4 m s^{-1} at Castilla, Spain. Dorman et al. (1995) also report standard deviations for cross-strait winds ranging from 2.1 m s^{-1} at Punta Cires to 4.3 m s^{-1} at Castilla. The observed wind speed variability can be compared to the NCEP–NCAR wind stress data using an approximate formula for conversion of wind speed to stress:

$$\tau = \rho_a C_{Da} u_a |u_a|, \tag{1}$$

where ρ_a is atmospheric density, approximately 1.25 kg m^{-3} near the sea surface; u_a is the 10-m atmospheric wind; and C_{Da} is an air–sea drag coefficient, approximately 1.3×10^{-3} for typical conditions (Kara et al. 2000). This leads to estimated NCEP–NCAR 10-m winds at 36°N, 6.5°W that have zonal and meridional standard deviations of, respectively, 7.5 and 5.3 m s^{-1}, which is about right for the zonal wind but somewhat higher than the observations for the meridional wind.

Some characteristics of the zonal τ_x and meridional τ_y wind stresses at 36°N, 6.5°W, 83 km west of Camarinal

TABLE 1. Summary of notation.

Symbol	Definition
η_{atl}	Sea level on the Atlantic side of Gibraltar Strait
η_{alb}	Sea level in the Alboran Sea
η_{med}	Sea level in the Mediterranean Sea
h	Sea level difference $\eta_{alb} - \eta_{atl}$
τ_x	Zonal surface wind stress at 36°N, 6.5°W
τ_y	Meridional surface wind stress at 36°N, 6.5°W
T_s	Time scale for the setup due to the along-strait wind
T_h	Delay due to Gibraltar Strait hydraulics
T_b	Barotropic equilibration time scale
T_f	Frictional adjustment time

Sill, and of the sea level time series from Fig. 2 are listed on Table 2. At the annual period, peak sea level occurs in late September–early October, which is consistent with the seasonal heating and cooling of the water column. The annual cycle amplitude of the Mediterranean and Alboran sea level is approximately twice that of Atlantic sea level: 7.4 versus 4.2 cm. The Alboran to Atlantic sea level difference h has an annual cycle amplitude of 3.3 cm with peak difference in early September. By comparison, the peak wind stress at the annual period occurs in mid-July for τ_x and early January for τ_y. This phase difference indicates that winds near Gibraltar are unlikely to be a primary driver of the Alboran to Atlantic sea level difference at the annual cycle. At the semiannual period, there is a much closer correspondence between the phase of h (137 days) and the phases of τ_x and τ_y (135 and 140 days, respectively). This indicates a possible dependence of h on τ_x and τ_y with an approximate transfer coefficient of 0.43 m Pa^{-1}.

To reduce the contamination of the sea surface height observations from processes other than barotropic wind-driven currents, for example, from the seasonal heating and cooling of the water column and from low-frequency changes in the hydraulic regime of Gibraltar Strait (Ross and Garrett 2000), the analysis that follows is restricted to periods ranging from 20 to 360 days, the 20-day cutoff being imposed in order to

match the 10-day TOPEX/Poseidon repeat cycle. Eighth-order Chebyshev type I filters are once again employed both for the low-pass and for the high-pass filters that are applied to the wind stress and sea level data. Table 2 shows that approximately 60% of the wind stress variance is at periods shorter than 20 days and that 25% of the wind stress variance is in the 20–360-day band. The remaining variance is at the annual cycle and longer periods.

Table 2 also shows that in all frequency bands, the Alboran and Mediterranean sea level variability is larger than that of the Atlantic Ocean near Gibraltar Strait. This is because Gibraltar Strait exchange processes have a disproportionately larger impact on Mediterranean rather than on Atlantic sea level variability given that the volume of the Mediterranean is approximately 4×10^{15} m^3 as compared with 1.4×10^{18} m^3 for the global ocean. In the 20–360-day band, the standard deviation of the Alboran–Atlantic sea level difference is 3.8 cm. The hypothesis investigated in this article is that some significant fraction of this variability results from the barotropic response to winds near Gibraltar. For this purpose, Fig. 3 compares the bandpassed wind stress time series with the bandpassed Alboran–Atlantic sea level difference. Black lines represent the zonal (τ_x; Fig. 3, top) and meridional (τ_y; Fig. 3, bottom) surface wind stresses at 36°N, 6.5°W and gray lines represent the sea level difference $h = \eta_{alb} - \eta_{atl}$. Note that given the coarse horizontal grid spacing of the NCEP–NCAR reanalysis (167 km zonally and 210 km meridionally near Gibraltar Strait), the wind stress at 36°N, 6.5°W represents a spatial average across the complete Gibraltar Strait and the adjoining coastal regions on the Atlantic side of the strait.

The correlation coefficient functions for the time series in Fig. 3 are displayed in Fig. 4 as a function of time lag. The thick dashed and solid lines represent, respectively, $\rho(h, \tau_x)$ and $\rho(h, \tau_y)$, the correlation of the observed Alboran–Atlantic sea level difference with zonal and meridional NCEP–NCAR wind stresses at 36°N,

TABLE 2. Summary characteristics of the sea level time series in Fig. 2 and of the surface wind stress from NCEP–NCAR reanalysis data at 36°N, 6.5°W. Units are as indicated, except for phase, which is time of peak amplitude in days from 1 January.

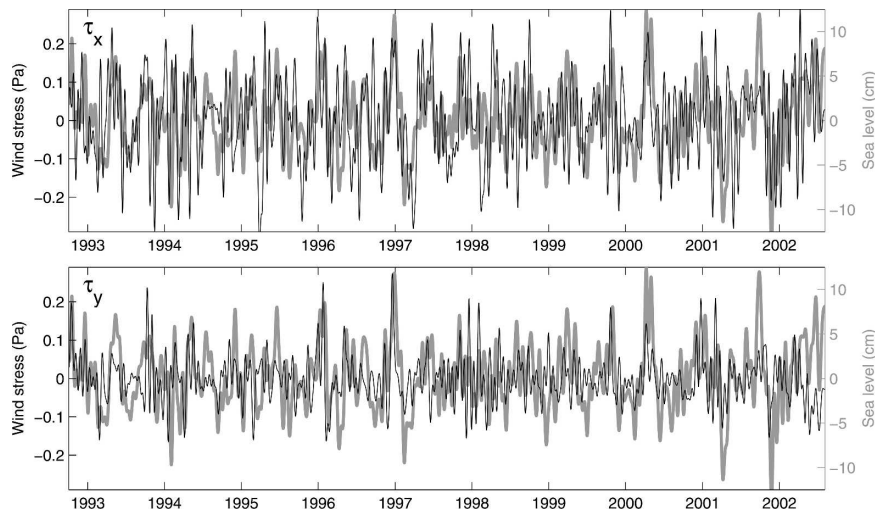|	η_{atl} (cm)	η_{alb} (cm)	η_{med} (cm)	h (cm)	τ_x (Pa)	τ_y (Pa)
Std dev	3.8	7.0	6.2	5.4	0.21	0.12
Std dev of 0–20-day band	—	—	—	—	0.16	0.09
Std dev of 20–360-day band	1.9	3.9	3.2	3.8	0.11	0.06
Semiannual amplitude	0.7	1.4	1.5	0.7	0.015	0.007
Semiannual phase in days	120	128	135	137	135	140
Annual amplitude	4.2	7.4	7.4	3.3	0.030	0.043
Annual phase in days	273	264	282	251	200	5

FIG. 3. Comparison of barometrically corrected TOPEX/Poseidon sea surface height and NCEP–NCAR wind stress near Gibraltar Strait for periods ranging from 20 to 360 days. Black lines represent (top) zonal (τ_x) and (bottom) meridional (τ_y) surface wind stresses at 36°N, 6.5°W, 83 km west of Camarinal Sill; positive values indicate eastward and northward wind stresses, respectively. Gray lines represent the sea level difference between the Alboran Sea and the Atlantic side of Gibraltar Strait; i.e., they are the high-passed difference $h = \eta_{alb} - \eta_{atl}$ from the bottom panel in Fig. 2.

6.5°W. The thin lines are the 95% confidence intervals computed by assuming that the joint distributions are binormal and that the decorrelation time scale is 20 days (Press et al. 1992). Sea level difference h exhibits significant correlation both with τ_x and with τ_y; h lags τ_x by 1 ± 5 days and τ_y by 5 ± 4 days. Confidence intervals for the time lags represent the regions above the peak correlation coefficient functions that are within the 95% confidence intervals.

The above analysis was repeated for different powers of wind stress, τ_x^p and τ_y^p with p ranging from 0.5 to 1.5. The power laws that maximize the correlation coefficients are $p = 0.85$ and $p = 1.05$ for, respectively, $\rho(h, \tau_x)$ and $\rho(h, \tau_y)$. But the differences in correlation coefficients are not significant at the 95% confidence interval; that is, there is not sufficient information to claim that p is different from 1.

By least squares fit, it is estimated that the ratio of the TOPEX/Poseidon sea level difference to the NCEP–NCAR surface wind stress near Gibraltar is

$$h(t) = (0.10 \pm 0.04)\tau_x(t - 1) + (0.19 \pm 0.08)\tau_y(t - 5),$$

(2)

where t is time in days, $h = \eta_{alb} - \eta_{atl}$ is the Alboran–Atlantic sea level difference in meters, and τ_x and τ_y are, respectively, zonal and meridional surface wind stresses in pascals at 36°N, 6.5°W; the 95% confidence intervals are estimated assuming no prior information

about the two coefficients, a decorrelation time scale of 20 days, and the assumption that data errors account for 75% of the observed variance (Wunsch 1996). Equation (2) explains 21% of the observed $\eta_{alb} - \eta_{atl}$ variance in the 20–360-day band, where the percent variance of time series x explained by time series y is defined as
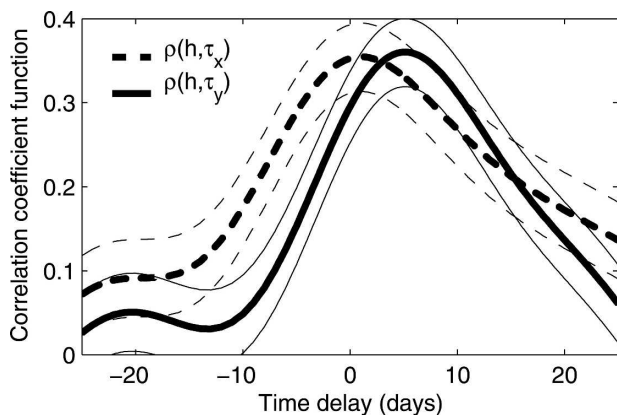

FIG. 4. Correlation coefficient function between the NCEP–NCAR surface wind stress near Gibraltar Strait and the barometrically corrected TOPEX/Poseidon sea level difference between the Alboran Sea and the Atlantic Ocean, $h = \eta_{alb} - \eta_{atl}$. Correlation functions are shown both for zonal, $\rho(h, \tau_x)$ is the thick dashed line, and meridional, $\rho(h, \tau_y)$ is the thick solid line, winds. Positive delay indicates that sea level lags wind stress. The thin lines are the 95% confidence intervals.

$$\text{Explained variance} = \frac{\text{var}(x) - \text{var}(x - y)}{\text{var}(x)} \times 100\%.$$

$$(3)$$

Note that the residual $x - y$ is not assumed to be un-correlated with x or y and therefore that the explained variance is not necessarily proportional with the square of the correlation coefficient.

3. Barotropic currents over the continental shelf

Having established that the barometrically corrected TOPEX/Poseidon sea level difference between the Atlantic and the Alboran Sea is correlated with NCEP–NCAR surface winds near Gibraltar, the next step is to determine whether this correlation is accidental or whether it indicates linear dependence. To establish a linear dependence, this section lays out a theoretical basis for the wind-driven Atlantic to Mediterranean sea level difference.

The dynamics of the barotropic currents over the continental shelf are discussed by Csanady (1974) using linearized equations of motion and a quadratic bottom friction law. When the wind direction is perpendicular to the coastline, the sea level response is often called barotropic or external wind setup (e.g., Gravili et al. 2001). When the wind direction is parallel to the coastline, the sea level response is that of a local coastal storm surge (e.g., Gill 1982, section 10.9). The wind setups due to cross-shore or alongshore winds are routinely observed by tidal stations near Gibraltar (e.g., García-Lafuente et al. 2004). The equations underlying these two processes are briefly reviewed below.

a. Setup due to along-strait wind

When the wind blows perpendicular to a coastline, a steady-state balance is quickly established between the surface wind stress and a pressure force due to the slope of the sea surface. The pressure force per unit volume is $\rho_w g \partial \eta / \partial x$, where ρ_w is the water density, g is the acceleration due to gravity, η is the sea surface height, and x is a coordinate perpendicular to the coastline. The wind stress force per unit volume is τ_x / H, where H is the depth of the water column and τ_x is the surface wind stress in a direction aligned with x. Thus, a force balance predicts the sea surface slope to be [Csanady 1982, Eq. (2.2)]

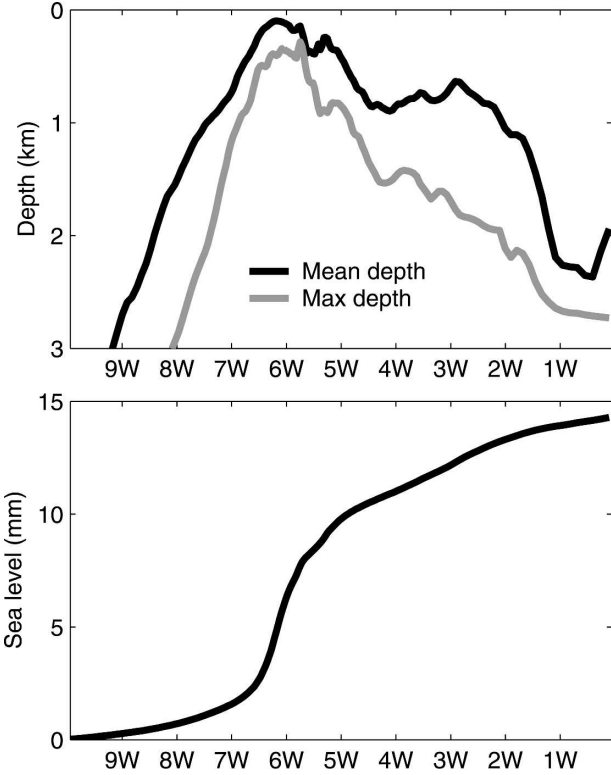$$\frac{\partial \eta}{\partial x} = \frac{\tau_x}{\rho_w g H}.$$

$$(4)$$

FIG. 5. (top) The meridionally averaged and maximum depth of Gibraltar Strait and the adjoining seas. The shallowest region is in the Atlantic, to the west of Camarinal Sill, between 6.5° and 5.75°W. (bottom) The sea level change due to Eq. (5) and to a steady along-strait surface wind stress of 0.1 Pa assuming a channel of constant width and an along-channel depth profile equal to the meridionally averaged depth profile.

Integrating, the magnitude of the steady-state barotropic wind setup is

$$\eta(x) = \frac{\tau_x}{\rho_w g} \int_0^x \frac{\xi}{H(\xi)} \, d\xi.$$

$$(5)$$

Equation (5) can also be used to estimate the along-strait sea level change due to along-strait winds above a channel of constant width and of depth $H(\xi)$ that varies only in the along-strait direction.

To give some idea of what the magnitude of the sea level response due to Eq. (5) would be in the absence of other physics, Fig. 5 shows an estimate of the sea level change assuming a steady zonal surface wind stress of $\tau_x = 0.1$ Pa and a channel of constant width that has an along-channel depth profile equal to the meridionally averaged depth profile of Gibraltar Strait and its adjoining seas. Notice that both the meridionally averaged (black line in Fig. 5, top) and the maximum (gray line in Fig. 5) depth profiles have their shallowest depths in a region immediately west of the Camarinal

Sill, between 6.5° and 5.75°W. This region is also where the maximum sea level change due to the along-strait wind setup is expected to occur (Fig. 5, bottom).

Ignoring any bathymetric inhomogeneities, hydraulic constraints within Gibraltar Strait, and the presence of the Mediterranean Basin, the sea level to wind stress equilibrium is expected to be reached for time scales longer than a setup time:

$$T_s = \frac{2L}{c}, \tag{6}$$

where $c = (gH)^{1/2}$ is the phase speed of shallow-water gravity waves and L and H are representative length and depth scales. For Gibraltar Strait and the adjoining shelf regions, from 7°W to 4.5°W, $L \approx 280$ km, $H \approx 730$ m, and $T_s < 2$ h.

b. Barotropic response of the Mediterranean Sea

Even though the setup time $T_s < 2$ h is extremely fast, the geostrophic or hydraulic controls (Garrett 2004) of the flow through the Strait of Gibraltar are expected to delay the Mediterranean sea level response to sudden changes in sea level on the Atlantic side of the strait. Theoretical (Bormans and Garrett 1989) and experimental (Garrett et al. 1989) results suggest that the ratio of the Gibraltar Strait throughflow to the sea level difference between the Atlantic and the Mediterranean is approximately $\alpha \approx 4 \times 10^7$ m^2 s^{-1}, that is, 0.4 Sv cm^{-1}. Therefore,

$$\frac{\partial h}{\partial t} = \frac{-\alpha h}{A_{\mathrm{med}}}, \tag{7}$$

where h is the Atlantic–Alboran sea level difference and $A_{\mathrm{med}} \approx 2.5 \times 10^{12}$ m^2 is the surface area of the Mediterranean Sea. Integrating Eq. (7) gives an exponential decay time of

$$T_h = \frac{A_{\mathrm{med}}}{\alpha} \approx 17 \text{ h}. \tag{8}$$

A delay may also be expected due to the time taken for a barotropic sea level change to establish itself within the Mediterranean Basin. This delay could be of order

$$T_b \approx \frac{2L_{\mathrm{med}}}{(gH_{\mathrm{med}})^{1/2}} \approx 18 \text{ h}, \tag{9}$$

where $L_{\mathrm{med}} \approx 3.9 \times 10^6$ m and $H_{\mathrm{med}} \approx 1.5 \times 10^3$ m are, respectively, the length and depth of the Mediterranean Sea.

c. Storm surge due to alongshore wind

As pointed out by Garrett (1983), there is considerable evidence to suggest that the alongshore wind component is much more important than the cross-shelf wind component in producing coastal sea level changes at low frequency. Consider a semi-infinite sea of uniform depth H bounded on the right by a straight boundary at $x = 0$. A wind stress parallel to the coast and of magnitude τ_y is applied starting at time $t = 0$. The wind forcing produces an Ekman current to the right of the wind in the Northern Hemisphere, which causes the surface to rise at a constant rate within a distance on the order of the barotropic Rossby radius of deformation from the coast. Ignoring friction, the rate of change of the sea surface elevation at the coast is [Gill 1982, Eq. (10.9.7)]

$$\eta(x, t) = \frac{\tau_y}{\rho_w c} t \exp(-x/a), \tag{10}$$

where x is the distance from the coastline, $a = c/f$ is the barotropic Rossby radius of deformation, and $f \approx 8.5 \times 10^{-5}$ s^{-1} is the Coriolis parameter. That is, the storm surge solution in Eq. (10) is proportional to the alongshore wind stress τ_y and increases linearly with time t. Equation (10) predicts a region of coastal influence that decays with $\exp(-x/a)$, where a is the radius of deformation based on some appropriate depth scale. Csanady (1974) suggests that, for typical coastal regions, the appropriate depth scale is the depth of the open ocean. A conservative estimate is $H = 284$ m, the depth of Camarinal Sill, which nevertheless results in $a \approx 620$ km, much longer than the 43-km width of Gibraltar Strait on the Atlantic side. Therefore, coastal processes on either side of the strait are expected to influence sea level at the entrance of the strait.

The storm surge solution due to alongshore wind stress also predicts a steadily increasing alongshore current, parallel to the coast, which is in geostrophic equilibrium with the pressure gradient due to Eq. (10). A steady state is reached when the bottom stress applied to this alongshore current is sufficiently large to balance the surface stress. Assuming (i) the existence of slow large-scale motions, that is, $\partial/\partial y \ll \partial/\partial x$ and $\omega \ll f$, where ω is the frequency; (ii) a shelf width smaller than the barotropic radius of deformation, which is the case here; and (iii) a bottom stress of the form $\rho_w C_D v|v|$, where C_D is a drag coefficient and v is the depth-averaged alongshore current, the steady-state solution is (Sandstrom 1980)

$$\frac{\partial \eta}{\partial x} = \frac{f}{g} \left(\frac{\tau_y}{\rho_w C_D} \right)^{1/2}. \tag{11}$$

Assuming a quadratic wind speed to wind stress conversion law as in Eq. (1), Eq. (11) predicts that the local storm surge sea level will scale linearly with wind speed, contrary to the wind setup solution in Eq. (4), which is expected to scale linearly with wind stress. Equation (11) together with Eq. (1) and a shelf width of 40 km was used by Garrett [(1983), Eq. (7.1)] to estimate that the Atlantic–Mediterranean sea level difference is approximately 1 cm for a 1 m s^{-1} wind.

The sea level to wind stress equilibrium is reached for time scales longer than a frictional adjustment time [Csanady 1982, Eq. 6.4b)]:

$$T_f = \frac{H}{2} \left(\frac{\rho_w}{C_D |\tau_y|} \right)^{1/2} . \tag{12}$$

The frictional adjustment time varies directly with depth and inversely with the square root of the wind stress and bottom drag coefficient. Below, it is argued that the relevant T_f for the Atlantic–Mediterranean sea level difference may be the time needed to establish the frictional equilibrium at the depth of Camarinal Sill: $H = 284$ m. A typical value for the nearshore bottom drag coefficient is $C_D = 0.002$ (e.g., Feddersen et al. 2003). For $\tau_y = 0.1$ Pa, the approximate frictional adjustment time is $T_f \approx 90$ h.

d. The role of Camarinal Sill

At steady state, with zero net transport through the Strait of Gibraltar, a pressure gradient $\partial \eta / \partial x$ normal to the Atlantic coastline is established via Eq. (11), causing an alongshore current,

$$v = \frac{g}{f} \frac{\partial \eta}{\partial x} , \tag{13}$$

in geostrophic balance with the pressure gradient. The North African coastline stretching from Cap Blanc, at the southern tip of the western Sahara, to Gibraltar exposes almost 2000 km of relatively straight, uninterrupted, and shallow continental shelf to south-southwesterly winds. For this reason, under steady homogeneous wind forcing, alongshore currents along the Moroccan coastline are expected to be larger than those in the Gulf of Cadiz or in the Alboran Sea. As the Moroccan coastal current reaches the opening of the strait, the assumptions used in deriving the storm surge equation [Eq. (11)] break down. The expectation is that, in an attempt to follow isobaths and hence to conserve potential vorticity (Csanady 1974), the Moroccan coastal current will be deflected toward the strait and accelerate as the isobaths get closer together. Some fraction of this coastal current, approximately that which follows isobaths deeper than 284 m, will be able to cross the opening of the strait at the Camarinal Sill and to join with (or to compete against) coastal currents on the north side of the Gibraltar Strait. The surface pressure gradient associated with this cross-strait current west of Camarinal Sill is expected to contribute to the sea level difference between the Atlantic and Mediterranean. The Moroccan coastal current that is inshore of the 284-m isobath continues into the Strait of Gibraltar, but it is to a large extent unable to cross the strait because of conservation of potential vorticity constraints. At steady state, some fraction of this current will be dissipated in shallow coastal regions while the remainder may contribute to additional changes in the Atlantic–Mediterranean sea level difference or be balanced by the outgoing current on the northern side of Gibraltar Strait.

To summarize, the Atlantic–Mediterranean sea level difference due to an alongshore wind-driven currents is expected to occur primarily offshore of the 284-m isobath on the Atlantic side of Gibraltar Strait west of Camarinal Sill. Given the irregular coastline and bathymetry of the strait, however, it is hard to estimate sea level at the 284-m isobath. With the hindsight of numerical simulation results, which are presented in section 4, sea level is set west of Tangier, near the closest approach of the 284-m isobath to the coastline, approximately 6 km offshore; the sea level amplitude at the 284-m isobath near Tangier is approximately 4 cm for a steady, homogeneous, south-southwesterly wind stress of 0.1 Pa.

4. Barotropic simulations of Gibraltar throughflow

The complicated bathymetry of Gibraltar Strait and the surrounding areas makes it difficult to obtain precise estimates on the impact of the along-strait wind setup and of alongshore windstorm surges on the Atlantic–Mediterranean sea level difference from analytic considerations alone. In this section, high-resolution barotropic simulations of the throughflow are used to test and to refine the analytic results.

Simulations are carried out using a global ocean configuration of the Massachusetts Institute of Technology general circulation model (MITgcm; Marshall et al. 1997a,b) with a telescoping horizontal grid that has increased resolution in the region near Gibraltar. Near Gibraltar, the horizontal grid spacing is ⅟₃₀°, approximately 3 km zonal by 3.7 km meridional. Grid spacing gradually increases to approximately 4° near the poles and in the Pacific Ocean. This grid configuration avoids having to make arbitrary decisions about where and what open boundary conditions to impose, it provides high horizontal resolution in the study region, and yet it
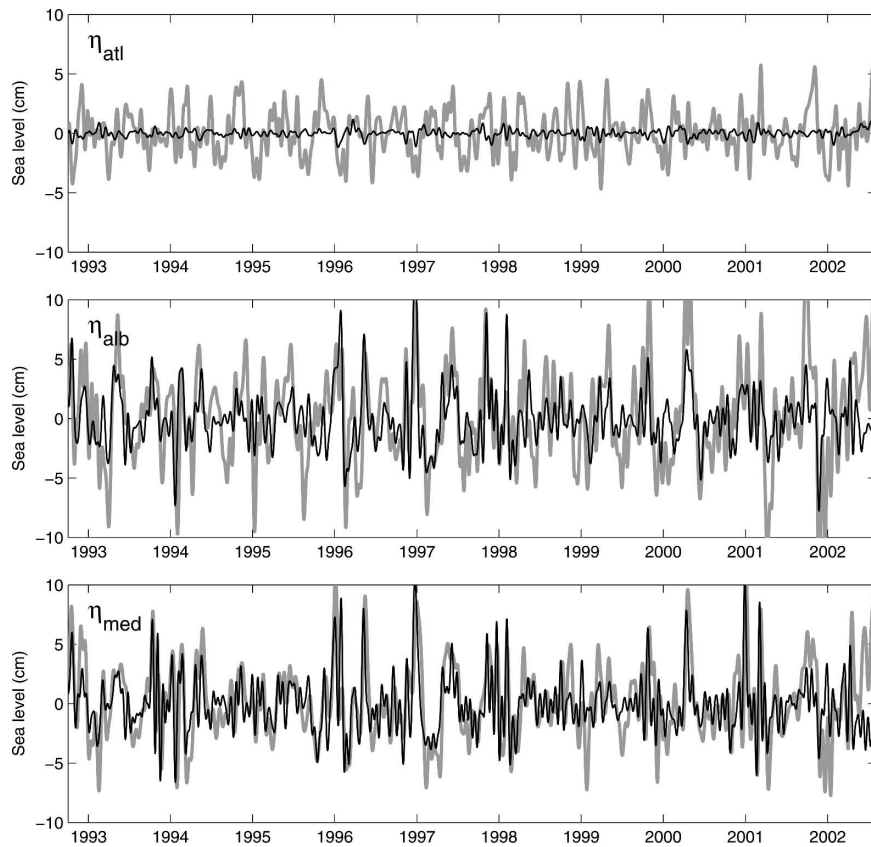
FIG. 6. Comparison of observed (gray lines) and simulated (black lines) sea levels for the 20–360-day band. (top) The mean sea level on the Atlantic side of Gibraltar Strait, (middle) the mean sea level in the Alboran Sea, and (bottom) the mean sea level in the Mediterranean, including the Alboran Sea.

results in a manageable 440×266 horizontal grid dimension.

Bathymetry is from the National Geophysical Data Center (NGDC) 2-min gridded global relief data ($2'$ Gridded Earth Topography; ETOPO2), which in the region of interest is obtained from the work of Smith and Sandwell (1997). The model's vertical grid extends to a maximum depth of 5200 m and comprises 15 vertical levels ranging in thickness from 50 m at the surface to 690 m at the bottom. The model employs the partial-cell formulation of Adcroft et al. (1997), which permits the accurate representation of the bathymetry despite the small number of vertical levels.

The model is barotropic in the sense that the potential temperature and salinity are kept constant throughout the domain and period of integration. Because of frictional processes and variable bathymetry, however, the horizontal velocity is not expected to be uniform throughout the water column.

The bottom friction is quadratic in the model's bottommost wet level with the drag coefficient $C_D = 0.002$.

Vertical viscosity above the bottom level is set to 1×10^{-3} m^2 s^{-1}. The horizontal viscosity is biharmonic and follows Leith (1996) but with the addition of a term proportional to the horizontal gradient of the divergence (B. Fox-Kemper 2005, personal communication). The model is initialized from rest and integrated using an implicit free-surface formulation and a time step of 90 s.

a. 1992–2003 integration

In a first numerical experiment, the barotropic model is integrated for the 1992–2003 period forced by the NCEP–NCAR surface wind stress. Figure 6 compares the barometrically corrected TOPEX/Poseidon observations with simulation results for the 20–360-day band. The model time series are constructed by sampling and averaging the simulated sea surface height in exactly the same way as the available data in order to minimize the differences that arise because of sampling. We have verified that the simulated time series sampled at the

TOPEX/Poseidon tracks are almost identical to the suitably filtered basin averages that are obtained from the full model fields. Therefore, aliasing due to the TOPEX/Poseidon sampling is likely to be small.

In the 20–360-day band, there is significant correlation between the observed and simulated time series of the mean sea level in the Alboran Sea ($\rho = 0.52$) and in the Mediterranean ($\rho = 0.56$) but not in the Atlantic Ocean ($\rho = 0.09$). The simulation explains 27% of the observed variance in the Alboran Sea, 28% of the observed variance in the Mediterranean, but only 0.2% of the observed variance on the Atlantic side of the Gibraltar Strait. This suggests that while the mean sea level on the Atlantic side of the Gibraltar Strait is dominated by internal variability (e.g., Ambar et al. 1999), which is not simulated by the barotropic model, the Alboran Sea and Mediterranean mean sea level variability has a large barotropic component in the 20–360-day band. The correlation coefficient between the simulated and observed Atlantic–Alboran sea level difference is $\rho = 0.55$ and the explained variance is 29%.

Repeating the data analysis carried out in section 2b, but replacing the TOPEX/Poseidon data with equivalently sampled model simulation results, the correlation coefficient functions between the simulated h and the NCEP–NCAR surface wind stress have peaks at $\rho(h, \tau_x) = 0.54$ and $\rho(h, \tau_y) = 0.62$ with delays of, respectively, 1 ± 2.5 days and 2.6 ± 2.5 days. The optimal power-law coefficients are, respectively, $p = 1.05$ and $p = 1.15$. Once again, there are no grounds to claim that these power laws are significantly different from 1. By using a least squares fit, it is estimated that the ratio of the simulated Atlantic–Alboran mean sea level difference to the NCEP–NCAR surface wind stress is

$$h(t) = (0.10 \pm 0.02)\,\tau_x(t-1) + (0.23 \pm 0.06)\,\tau_y(t-2.6),$$

(14)

which is consistent with Eq. (2), which was derived using the TOPEX/Poseidon mean sea level data. Equation (14) explains 58% of the variance of the simulated h in the 20 to 360-day band.

Last, there is a high degree of similarity between the Alboran and Mediterranean mean sea level time series. For barometrically corrected TOPEX/Poseidon data, the Alboran mean sea level explains 21% of the Mediterranean mean sea level variance and the correlation coefficient between the two time series is $\rho = 0.70$. For the simulation, the explained variance is 59% and the correlation coefficient is $\rho = 0.80$. The similarity of the Alboran and Mediterranean mean sea level time series provides a pathway for winds near Gibraltar Strait to affect the mean sea level of the entire Mediterranean.

b. Response to steady wind forcing

To better understand the transient and steady-state responses to the wind forcing, the barotropic model is initialized from rest and forced with steady zonal and meridional winds for 40 days. A uniform surface wind stress is applied on a region with a 500-km radius, centered near Gibraltar (36°N, 6°W). Outside this region, the wind stress decays gradually with a sinusoidal profile. The complete footprint of the wind stress perturbation has a radius of 4000 km; therefore, it includes all of the northwestern African coastline, where alongshore windstorm surges develop that are expected to influence the Atlantic–Mediterranean sea level difference.

Figure 7a shows the steady-state sea level anomaly near Gibraltar Strait from an integration forced by zonal wind $\tau_x = 0.1$ Pa. The corresponding meridionally averaged sea level as a function of longitude is shown by the blue line in Fig. 7c. The Atlantic to Mediterranean sea level rise is approximately 3.6 cm, most of it (≈ 2.3 cm) occurring between 7° and 6°W on the Atlantic side of the strait. This sea level rise is larger than can be explained by the along-strait wind setup alone, which from Eq. (5) and Fig. 5 is estimated to be approximately 1.4 cm. The increased coastal sea level along Morocco and the decreased coastal sea level along the Iberian Peninsula are consistent with the storm surge solution due to the alongshore component of the wind.

The blue line in Fig. 7e is a time series of the meridionally averaged sea level difference between 7.5° and 5.5°W. As predicted by Eq. (6), the spinup for the along-strait wind begins with an extremely fast adjustment, order 2 h, which initially causes the sea level on the Atlantic side to be higher by approximately 0.6 cm than that of the Mediterranean. This is because to a first approximation the Gibraltar Strait initially acts as a closed boundary. On the Atlantic side, the setup due to cross-shelf winds raises the sea level while on the Alboran side it lowers the sea level. The Mediterranean sea level responds to the Atlantic sea level change with a delay of approximately 36 h, which is consistent with Eqs. (8) and (9). The overall delay is of order 79 h, comparable to the frictional adjustment time [Eq. (12)] of a storm surge due to the alongshore wind.

Figure 7b shows the steady-state sea level from an integration forced by the meridional wind $\tau_y = 0.1$ Pa. The Mediterranean–Atlantic sea level rise is of order 4.6 cm, with most of this rise (≈ 3.3 cm) once again occurring on the Atlantic side of the strait. For the
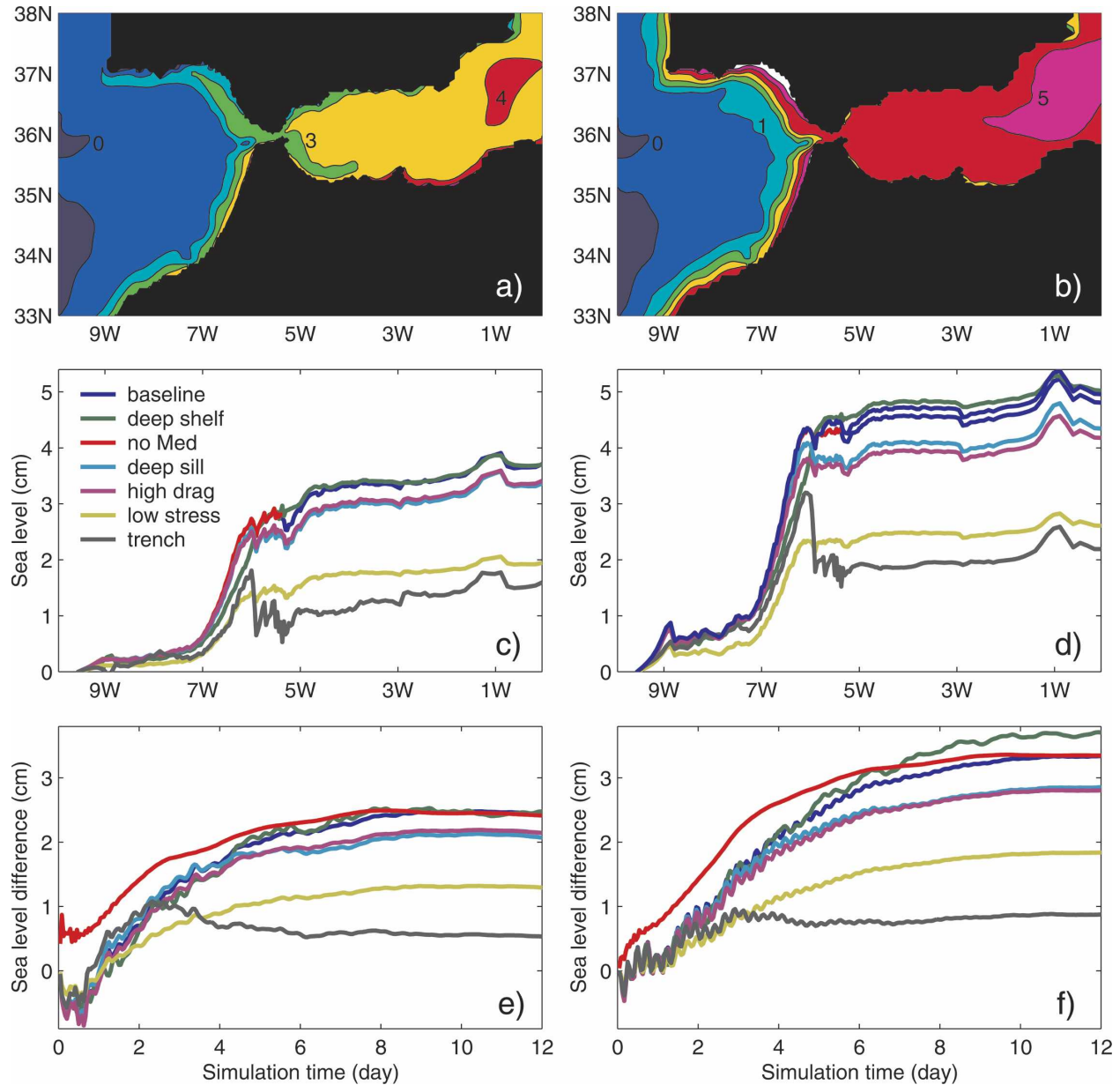
FIG. 7. Simulated sea level change from the Atlantic to the Mediterranean for (left) zonal and (right) meridional wind forcings. (a) Steady-state (day 40) response of the baseline configuration to a zonal wind stress $\tau_x = 0.1$ Pa. Contour intervals are 1 cm. (b) Same as in (a) but for meridional wind stress $\tau_y = 0.1$ Pa. (c) Meridionally averaged steady-state sea level for the region shown in (a). Results are from the baseline integration and from six sensitivity studies as indicated in the legend. (d) Same as in (c) but for the meridional wind stress. (e) Time series of the meridionally averaged sea level difference between 7.5° and 5.5°W. Legend is as in (c). (f) Same as in (e) but for the meridional wind stress. Note that time series in (e) and (f) exclude the sea level rise on the Atlantic side of Gibraltar Strait so that they can be compared with the "no Mediterranean" sensitivity experiments.

meridional wind, the alongshore wind component on the Atlantic side of Gibraltar Strait is much larger than the along-strait wind component. Therefore, the response is expected to be primarily due to the storm surge solution in Eq. (11). Notice that to a first approximation, the Atlantic–Mediterranean sea level difference occurs west of Tangier, offshore from the 284-m

isobath. The delay of the sea level response is approximately 98 h, comparable to the theoretical prediction $T_f \approx 90$ h for $\tau_y = 0.1$ Pa from (12).

c. Sensitivity experiments with steady wind forcing

To test the scaling predictions made in section 3, the results from 12 additional steady-state sensitivity ex-

TABLE 3. Summary of sea level difference and time delay for 12 numerical experiments forced by steady winds near Gibraltar Strait. Sea level difference is reported between 7.5° and 5.5°W. Delay is defined as the time taken for the sea level difference to reach 63% of its steady-state value. The results presented in this table exclude the sea level rise on the Atlantic side of Gibraltar Strait so that they can be compared with the "no Mediterranean" sensitivity experiments.

Expt	Forcing (Pa)	Sea level (cm)	Delay (h)
Baseline	$\tau_x = 0.1$	2.4	79
	$\tau_y = 0.1$	3.3	98
Low stress	$\tau_x = 0.05$	1.3	79
	$\tau_y = 0.05$	1.8	95
High drag	$\tau_x = 0.1$	2.1	78
	$\tau_y = 0.1$	2.8	90
Deep sill	$\tau_x = 0.1$	2.0	79
	$\tau_y = 0.1$	2.8	82
No Mediterranean	$\tau_x = 0.1$	2.4	52
	$\tau_y = 0.1$	3.3	68
Deep shelf	$\tau_x = 0.1$	2.4	97
	$\tau_y = 0.1$	3.7	111
Trench	$\tau_x = 0.1$	0.5	24
	$\tau_y = 0.1$	0.9	48

periments, 6 with zonal wind forcing and 6 with meridional wind forcing, are compared with the two baseline solutions in Fig. 7 and Table 3.

1) LOW STRESS

In a first set of experiments, the wind stress is reduced from 0.1 to 0.05 Pa. Relative to the baseline experiments, the sea level difference between 7.5° and 5.5°W decreases from 2.4 to 1.3 for the zonal and from 3.3 to 1.8 for the meridional forcing. The ratio of the sea level difference between the baseline and the low-stress experiments is approximately 1.8, which is smaller than the predicted ratio of 2 for the wind setup solution in Eq. (4) and larger than the predicted ratio of $2^{1/2}$ for the storm surge solution from Eq. (11). The expectation, especially for the meridional experiments, is that the sea level response is dominated by the storm surge solution. Therefore, the ratio should be closer to $2^{1/2}$ than to 2. The time delay between the wind forcing and the sea level difference is approximately the same for the low-stress experiments as it is for the baseline experiments (see Table 3). This is contrary to Eq. (12), which predicts that the frictional adjustment time is inversely proportional to the square root of the wind stress.

Csanady (1974) provides a possible explanation for the above discrepancies by noting that as depth is increased, the assumption of the bottom stress of form $\rho_w C_D v|v|$, which is used to derive Eqs. (11) and (12), breaks down. In the numerical simulations, this as-

sumption breaks down for depths greater than 50 m, the thickness of the top level. Above the bottommost wet level, the simulation employs a constant vertical viscosity of 1×10^{-3} m^2 s^{-1}. Therefore, the effective C_D acting on the depth-averaged v would tend to be larger as the wind stress τ_y is decreased. As a consequence, for small τ_y, the sea level and the frictional time scale from the numerical simulations will tend to be smaller than the analytic predictions in Eqs. (11) and (12) and the power-law dependence of the sea level difference to wind stress would tend to be closer to 1 than to ½.

2) HIGH DRAG

In a second set of experiments, the bottom drag coefficient C_D is doubled to a value of 0.004 as compared with a value of 0.002 in the baseline integration. A doubling of the bottom drag coefficient is not expected to impact the wind setup solution [Eqs. (4) and (6)], but it is expected to reduce the amplitude [Eq. (11)] and the delay [Eq. (12)] of the storm surge solution. The ratios between the meridional baseline and the high-drag experiments for the sea level difference and the delay are, respectively, 1.2 and 1.1, which are lower than the expected value of $2^{1/2}$. Once again, the proposed explanation for this discrepancy is that C_D in the numerical simulations acts upon the bottom layer velocity only, while Eqs. (11) and (12) were derived assuming that C_D acts upon the depth-averaged velocity.

3) DEEP SILL

In a third set of experiments, the minimum depth of Gibraltar Strait is increased by 50% from 284 to 426 m. This leads to a decrease in the fraction of the Moroccan coastal current that is able to cross Gibraltar Strait west of Camarinal Sill. It also leads to a decrease in the sea level difference comparable to that of the high-drag experiment, which is consistent with the hypothesis that the wind-driven Atlantic–Mediterranean sea level difference is in part controlled by the depth of Camarinal Sill.

4) NO MEDITERRANEAN

In a fourth set of numerical experiments, Gibraltar Strait is blocked at 5.3°W and the Mediterranean Sea is removed. As expected, the presence or absence of the Mediterranean Sea has no steady-state impact on the Gibraltar Strait sea level. It does however impact the transient solution, as is seen from the differences between the red and blue lines in Figs. 7e and 7f.

First, the wind setup response to the zonal forcing does not cause an initial negative sea level response, as is the case for all of the experiments that include a

Mediterranean Basin (see Fig. 7e). Instead, an initial positive sea level response of approximately 6 mm is obtained. The expectation is that this fast response is dominated by the setup in Eq. (5) and the time scale from Eq. (6). The 6-mm estimated sea level rise between 7.5° and 5.5°W is very close to that estimated for a channel of constant width that has an along-channel depth profile equal to the meridionally averaged depth of Gibraltar Strait and the adjoining regions (Fig. 5). It is therefore estimated that for a steady zonal forcing of 0.1 Pa, the setup due to the along-strait winds contributes approximately 25% of the Atlantic–Mediterranean sea level change between 7.5° and 5.5°W.

Second, the time delay is reduced by approximately 30 h relative to the baseline integrations (see Table 3). This delay reduction corresponds to the time it takes to change the mean Mediterranean sea level following a sea level disturbance on the western side of Gibraltar Strait and it is comparable to the theoretical expectation of $T_h + T_b \approx 35$ h estimated from Eqs. (8) and (9).

5) Deep shelf

In a fifth set of experiments, the Atlantic and Alboran continental shelves are lowered to 284 m, the depth of Camarinal Sill. This deepening is expected to decrease the setup due to the along-strait wind [Eq. (4)] but to have a relatively small impact on the storm surge due to the alongshore wind [Eqs. (11) and (12)]. Instead, the impact of a deeper shelf on the experiment forced with the meridional wind is to increase the time delay and the steady-state sea level difference by approximately 12% relative to the baseline experiment. The larger sea level difference occurs because the deepening of the shelf provides a less convoluted path for the Moroccan coastal current to cross the opening of Gibraltar Strait. The experiment forced by the zonal wind shows very little change relative to the baseline. We speculate that this is because the decrease in the along-strait setup response due to the deeper shelves is compensated for by an increase in the alongshore storm surge response.

6) Trench

In a sixth set of sensitivity experiments, the model bathymetry includes a deep and narrow trench, 2 km deep and 3.7 km (one model grid) wide, passing through the middle of Gibraltar Strait from the Atlantic to the Mediterranean. The presence of this trench interrupts the Moroccan coast–Gulf of Cadiz current, resulting in a much smaller Atlantic–Mediterranean sea level difference than any of the other experiments. Figure 8 shows depth-averaged currents near Gibraltar Strait for the "baseline" and for the "trench" experiments.

The baseline experiments show that the Moroccan coastal current is accelerated as it nears Tangier, that it is deflected toward Gibraltar Strait, and that a fraction of this current is able to cross the strait immediately west of Camarinal Sill near the 284-m isobath. The fraction of the Moroccan coastal current that is inshore of the 284-m isobath continues into the Strait of Gibraltar where it is in part dissipated in shallow coastal regions. Notice that the alongshore currents are also established along the Iberian coastline, southward flowing for zonal forcing and northward flowing for meridional forcing, as expected. However, these currents are much weaker than their Moroccan counterparts because of the limited fetch of the Iberian continental shelf; hence, the Iberian coastal currents offer little resistance or amplification to the fraction of the Moroccan coastal currents that cross the strait near the 284-m isobath. The surface pressure gradient associated with the portion of the Moroccan current that crosses Gibraltar Strait contributes approximately 1.5 cm of sea level rise to the zonal wind experiment (Fig. 8a) and 2.5 cm to the meridional wind experiment (Fig. 8b).

In the trench experiments (Figs. 8c and 8d), the same alongshore currents as in baseline experiments are established. But contrary to the setup of the baseline experiments, the Moroccan coastal currents are to a large extent unable to cross Gibraltar Strait. Instead, they are mostly deflected through the strait, where they are in part dissipated in shallow coastal regions and in part balanced by outflowing water inside the trench and on the northern side of the strait. Therefore, while these currents affect the coastal sea level along the southern coast of Gibraltar Strait, they have a much smaller impact on the Atlantic–Mediterranean sea level difference than they do in the baseline experiments.

The net Atlantic to Mediterranean sea level rise in the trench experiments is approximately 45% of that in the baseline experiments (Figs. 7c and 7d). For the experiment forced by the zonal wind, this may be due in part to the along-strait wind setup. In particular, the net sea level change of the zonal wind trench experiment (black line in Fig. 7c) is very similar to the theoretical estimate for the wind setup due to the along-strait wind in a constant-width channel. The wind setup corresponding to Fig. 5, but with an average depth including the trench, predicts a 1.2-cm sea level difference from 10°W to 0° as compared to 1.3 cm in the numerical simulation. For the meridional wind trench experiment, the Atlantic–Mediterranean sea level difference may be due in part to the fraction of the Moroccan coastal
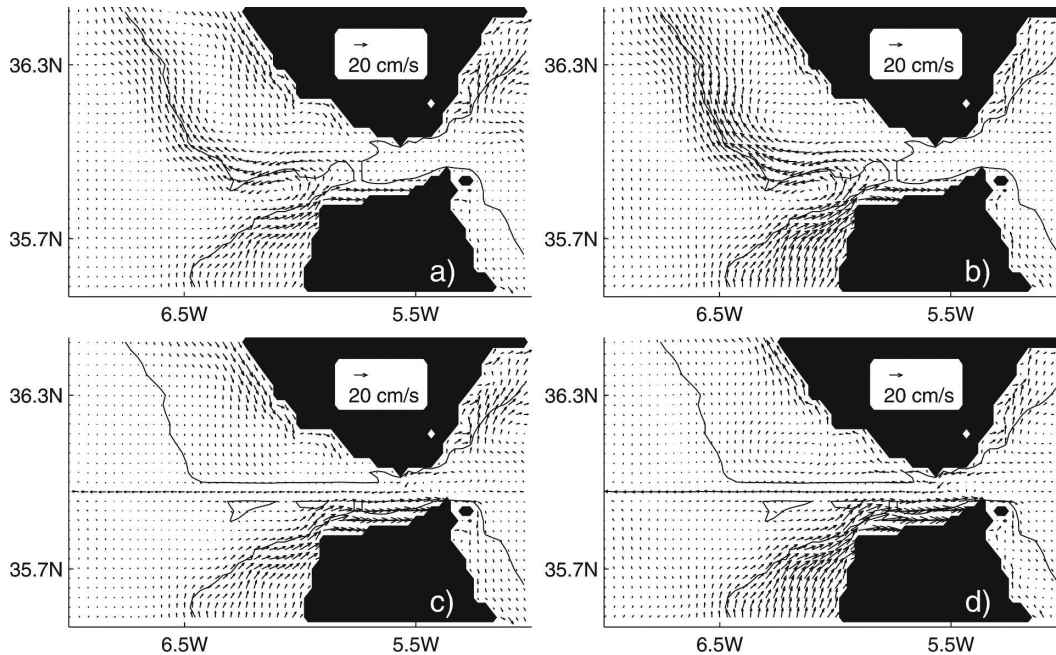
FIG. 8. Depth-averaged currents near Gibraltar Strait for the "baseline" and for the "trench" experiments. (a), (b) Currents for the baseline experiments, which have realistic bathymetry. (c), (d) The trench experiments, which have a 2-km-deep and 3.7-km-wide trench in the middle of Gibraltar Strait. Both (a) and (c) are forced by steady zonal wind $\tau_x = 0.1$ Pa and (b) and (d) are forced by steady meridional wind $\tau_y = 0.1$ Pa. The thin solid lines represent the 285-m isobath in the model bathymetry. The length of the vectors is proportional to the current speed and not to the vertical transport.

current that is not dissipated by frictional processes within Gibraltar Strait.

d. Sensitivity experiments with cyclical wind forcing

To further investigate the barotropic model's response, two final sets of numerical experiments have been conducted. A first set of experiments is forced with spatially homogeneous but time-varying zonal wind stress. The second set of experiments is forced with spatially homogeneous but time-varying meridional wind stress. In both sets of experiments the temporal variation is sinusoidal with periods ranging from 6 h to 28 days and an amplitude of 0.1 Pa. Figure 9 documents some of the results from these two sets of experiments.

At periods longer than 20 days, the transfer function of the sea level difference between 8.5° and 1.5°W to the surface wind stress asymptotes to 0.35 m Pa^{-1} with a time lag of 3.5 days and to 0.4 m Pa^{-1} with a time lag of 4.5 days for, respectively, the zonal and meridional wind forcing. These values are close to the steady-state model response (the blue lines in Figs. 7e and 7f).

At short periods, the model's response is more complicated. For zonal forcing, there is a peak in the transfer function at the 30-h period and a sharp phase tran-

sition at the 60-h period. The latter corresponds to the transition from a negative to positive sea level anomaly in the experiment forced by the steady zonal wind (blue line in Fig. 7e). For meridional forcing, there are peaks at the 6-, 24-, and 54-h periods. A possible rectification of the high-frequency response, that is, a small bias in the sea level response to high-frequency winds, is noted but not investigated.

5. Summary and concluding remarks

Evidence presented here supports the hypothesis of Garrett (1983), García-Lafuente et al. (2002b), and Fukumori et al. (2007) that a significant fraction of the Atlantic–Mediterranean sea level difference variability, excluding atmospheric pressure and seasonal cycle contributions, is a barotropic response to winds in the vicinity of Gibraltar Strait. A high-resolution barotropic numerical simulation forced by NCEP–NCAR reanalysis surface wind stress for the years 1992–2003 has a Mediterranean mean sea level variability in the 20–360-day band that is significantly correlated with barometrically corrected TOPEX/Poseidon data ($\rho = 0.56$) and that explains 28% of the observed variance. This simulation suggests that the wind-driven barotropic variabil-
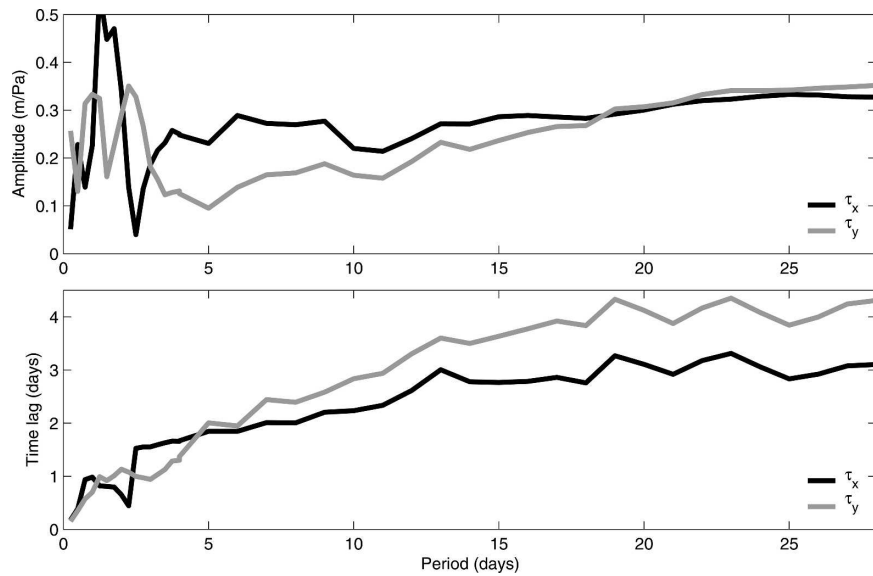
FIG. 9. The simulated sea level change between 8.5° and 1.5°W for the zonal (τ_x) and meridional (τ_y) wind forcings. The model is forced with a spatially homogeneous wind stress that varies sinusoidally in time with amplitude 0.1 Pa and periods ranging from 6 h to 28 days. (top) The amplitude of the wind stress to the sea level transfer function in m Pa^{-1}. (bottom) The time lag of the sea level response in days.

ity has a maximum amplitude of 20-cm peak to peak and a standard deviation of 3 cm for periods ranging from days to years. The residual variance of the Mediterranean mean sea level time series is attributed to hydraulic effects in Gibraltar Strait and to mass and buoyancy fluxes that are not represented in the barotropic simulation, to residual errors in the barometric and tidal corrections, and to various other model and data errors.

Theoretical arguments and numerical experiments establish that this wind-driven sea level difference between the Atlantic and the Mediterranean is in part caused by storm surges due to alongshore winds along the North African coastline on the Atlantic side of Gibraltar Strait. A fraction of the Moroccan coastal current, offshore from the 284-m isobath, is deflected across Gibraltar Strait. The geostrophic surface pressure gradient associated with this cross-strait current west of Camarinal Sill plays a key role in establishing Atlantic–Mediterranean sea level difference at the stationary limit.

For periods greater than a few days, the steady-state storm surge solution in Eq. (11) predicts that the sea level response is proportional to the square root of the wind stress and to the width of the North African continental shelf and that it is inversely proportional to the square root of the bottom drag coefficient. Observations and numerical simulations, however, suggest that

the dependence of sea level on the wind stress may be closer to linear.

The Atlantic–Mediterranean sea level difference is also controlled in part by the along-strait wind setup [Eq. (4)], with a contribution proportional to the along-strait wind stress and to the length of Gibraltar Strait and the adjoining regions and inversely proportional to the depth. For the zonal component of the wind, the wind setup is expected to account for approximately 25% of the Atlantic–Mediterranean sea level change between 7.5° and 5.5°W; an additional along-strait wind setup contribution, possibly up to 40% of the total, may come from the Alboran Sea (cf. Fig. 5, bottom panel, to the baseline experiment in Fig. 7c). Adding the storm surge and wind setup contributions, and given the orientation of the North African shelf on the Atlantic side of Gibraltar Strait, the Mediterranean sea level is expected to be highest relative to the Atlantic for steady south-southwesterly winds and lowest for north-northeasterly winds.

In the 20–360-day band, the average transfer coefficients between the Atlantic–Alboran sea level difference and the surface wind stress at 36°N, 6.5°W, estimated from barometrically corrected TOPEX/ Poseidon data and the NCEP–NCAR reanalysis, are 0.10 ± 0.04 m Pa^{-1} with 1 ± 5 day lag and 0.19 ± 0.08 m Pa^{-1} with 5 ± 4 day lag for, respectively, the zonal and meridional wind stresses. Equivalent estimates de-

rived from a high-resolution barotropic simulation driven by the NCEP–NCAR reanalysis surface wind stress are 0.10 ± 0.02 m Pa^{-1} with 1 ± 2.5 day lag and 0.23 ± 0.06 m Pa^{-1} with 2.6 ± 2.5 day lag for the zonal and meridional winds, respectively.

The above data and model-derived transfer functions are consistent with each other to within their respective uncertainties, but they differ from the numerical experiments with spatially homogeneous surface forcing (Figs. 7–9). The transfer function amplitudes estimated from the steady wind experiments are approximately twice those estimated from data or from numerical simulations with realistic forcing. We speculate, but have not investigated in detail, that this difference is caused by spatial inhomogeneities in the winds. Specifically, the storm surge amplitude near Gibraltar Strait depends on the total fetch of the south-southwesterly winds on the North African coastline.

Although the full physics have not yet been established, the numerical experiments and observations presented herein suggest a central role for Camarinal Sill and for the alongshore winds and currents along the North African coastline near the western entrance of Gibraltar Strait in causing wind-driven sea level difference between the Atlantic and the Mediterranean. A substantial contribution, approximately 25%–40% for the zonal wind component, is also expected from along-strait wind setup.

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