Inflow interruption by meteorological forcing in the Strait of Gibraltar

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[1] A case of extreme meteorologically forced fluctuation of net flow through the Strait of Gibraltar is analysed. The Atlantic water inflow was interrupted during some days and the net flow reached a peak of −1.5 Sv towards the Atlantic Ocean. In spite of the rapid increase of atmospheric pressure that triggered this episode, the expected inflow contribution to the net flow induced by pressure variation should not exceed 75% of the mean inflow, insufficient to reverse the inflow. Wind stress acting on the upper layer can induce important inflow and therefore net flow fluctuations. A simple model is proposed in which wind stress intensity as a function of frequency determines the inflow response. For low frequency fluctuations and moderate wind speeds the model predicts a gain that, if added to the atmospheric pressure effect, could bring the inflow fluctuation peak beyond the mean, thus explaining the inflow interruption. *INDEX TERMS:* 4504 Oceanography: Physical: Air/sea interactions (0312); 4512 Oceanography: Physical: Currents; 4243 Oceanography: General: Marginal and semienclosed seas; 4564 Oceanography: Physical: Tsunamis and storm surges. **Citation:** García Lafuente, J., J. Delgado, and F. Criado, Inflow interruption by meteorological forcing in the Strait of Gibraltar, *Geophys. Res. Lett.*, *29*(19), 1914, doi:10.1029/2002GL015446, 2002.

1. Introduction

[2] Evaporative and buoyancy losses in the Mediterranean Sea are responsible for the exchange flow through the Strait of Gibraltar (see Figure 1). Yearly average flow towards the Mediterranean Sea (Q_1 or inflow hereinafter, positive) and towards the Atlantic Ocean (Q_2 or outflow hereinafter, negative) are around 0.8 Sv [*Bryden et al.*, 1994; *Tsimplis and Bryden*, 2000; *Bascheck et al.*, 2001], with interannual fluctuations that are not yet well known. The net flow ($Q_0 = Q_1 + Q_2$) needed to compensate for evaporation in the Mediterranean Sea is around 0.05 Sv, one order of magnitude smaller than Q_1 or Q_2, but it undergoes fluctuations that are many times this yearly-averaged value. The paradigm is a tidal flow with an M_2 component of almost 3 Sv amplitude [*García Lafuente et al.*, 2000].

[3] It has long been known that meteorologically induced subinertial fluctuations of Q_0 are mainly forced by the variable atmospheric pressure over the Mediterranean [*Crepon*, 1965; *Garrett*, 1983; *Candela et al.*, 1989]. A barotropic numerical model of the Mediterranean Sea and North Atlantic Ocean forced by atmospheric pressure and diagnosed wind stress over the domain, presented in *García Lafuente et al.* [2002a], confirmed that most of Q_0 variance was induced by atmospheric pressure over the Mediterranean. However, they also suggested that wind stress has increasing importance during some wind events.

[4] The response of Q_1 and Q_2 to externally imposed atmospheric forcing, an important issue regarding the circulation variability of the adjacent basins, is better addressed carrying out the decomposition $Q_k(t) = Q_{mk} + q_k(t)$, $k = 0,1,2$ for net flow, inflow and outflow, respectively, m indicating the yearly averaged value. In this work, the fluctuating part, $q_k(t)$, is the atmospherically forced contribution. Obviously, $q_0(t) = q_1(t) + q_2(t)$, $q_1(t)$ and $q_2(t)$ showing remarkable tendency toward the same sign, either positive or negative, suggesting a basically barotropic response [*García Lafuente et al.*, 2002a]. These authors also showed that, on average, q_1 (q_2) was 60% (40%) of q_0. On the other hand, the simple model by *Candela et al.* [1989], whose predictions are fairly good, yields a gain of around 0.08 Sv/mb for q_0 in the meteorological frequency band ($\sim$0.08 to 0.3 cpd). This in turn would give gains of somewhat less than 0.05 Sv/mb and somewhat more than 0.03 Sv/mb for q_1 and q_2, respectively.

[5] Typical fluctuations in the spatially averaged atmospheric pressure over the Mediterranean Sea have a standard deviation of about 5 mb (see panel h of Figure 2), which would induce inflow and outflow fluctuations of 0.2 to 0.3 Sv, too small to produce flow reversals. Even wintertime strong atmospheric pressure fluctuations of twice the standard deviation do not appear to be sufficient to surpass the mean values Q_{m1} or Q_{m2}. Therefore, atmospherically forced flow reversals in the Strait of Gibraltar are remarkable events, which cannot be explained by atmospheric pressure changes alone. This paper analyses one such event observed in early 1998 during Canary Islands Azores Gibraltar Observations (CANIGO) project.

2. Data and Methods

[6] Current meter data from October 26, 1997 to March 30, 1998 were collected at the eastern part of the Strait of Gibraltar (see Figure 1) at 40m, 70m, 140m, 200m and 550m depth. The tidal contribution was removed from the observations as explained in *García Lafuente et al.* [2000]. The filtered time series, presented in the five top panels of Figure 2, were used to estimate inflow, outflow and the net flow. The isohaline S = 37.77 PSU that maximised the exchanged flows during that period was used as the interface. This result, as well as more details about flow computation, can be seen in *García Lafuente et al.* [2002a]. The final time series of flows are plotted in panels (f) and (g) of Figure 2.

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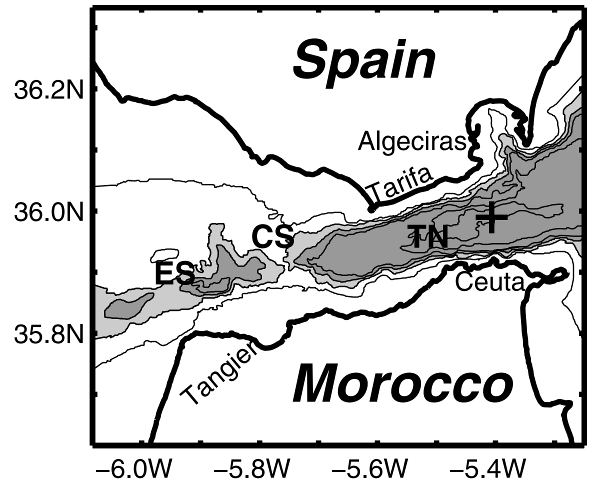

Figure 1. Map of the Strait of Gibraltar. Isobath depths are 100m, 200m, 290m, 400m, 700m and 900m. Depths greater than (290m) have been shaded.

[7] Wind speed and direction and atmospheric pressure in Tarifa and Ceuta (see Figure 1) were collected from Instituto Nacional de Meteorolgía (INM), Spain. Atmospheric pressure and wind stress fields over the Mediterranean Sea and North Atlantic Ocean generated through the High Resolution Limited Area Model (HIRLAM) by INM with half degree spatial resolution were also available.

[8] Sea level heights in Ceuta and Algeciras ports in the Strait were collected from Instituto Español de Oceanografía in order to estimate the across-strait sea level difference. The good visual correlation between this difference and the along-strait velocity shown in Figure 2a (correlation coefficient r = 0.78) ensures that geostrophy is satisfactorily verified.

3. The Event of Inflow Reversal

[9] All these data have been used to illustrate the inflow reversal that happened during early February 1998 (shaded rectangle in Figure 2) and to investigate its causes. The uppermost current meter registered westward currents during those dates. It was so unusual that further confirmation of the event was searched in the across-strait sea level difference, which is an independent measurement of the horizontally integrated along-strait velocity. The change of sign in sea level difference during the event confirms the reality of the inflow reversal.

[10] The primary cause for this event was the very large increase of atmospheric pressure over the Mediterranean Sea that took place from 5 to 11 of February (Figure 2h). All instruments were sensitive to this change. The inflow decrease induced by this atmospheric pressure change, which is characterised by a 10 mb amplitude and a period of 10 days ($\sim$0.1 cpd of frequency) in Figure 2h, would be of 0.5 to 0.6 Sv according to our previous discussion. This is significantly less than 0.8 Sv, the amount necessary to reverse the inflow. Moreover, the expected induced net flow should be around -0.8 Sv, whereas Figure 2f shows that $q_0(t)$ exceeded -1.5 Sv. Most of it came from the inflow change, which was more sensitive than the outflow (Figure 2g). Strong easterlies in the Strait blowing simultaneously with the atmospheric pressure change (Figure 2i) appear to be responsible for the enhanced response of the inflow and net flow.

4. Wind Forcing

[11] Two questions concerning wind-induced inflow variations are addressed: the spatial extension of the force, and the complex gain of inflow to wind forcing. Panels in Figure 3 show the time lagged covariance between the inflow time series and the series of wind stress at the grid points of a 0.5 × 0.5 mesh covering the area 25°W to 15°E and 30°N to 45°N. Vectors have been constructed taking the covariance of the inflow to the *x* wind stress and to the *y* wind stress as the *x* and *y* components, respectively. In order to plot representative values, a significance test has been performed by computing the standard deviation of the modulus of the covariance vector at each lag and selecting those vectors that exceeded the mean covariance by 1.5 times the standard deviation. The

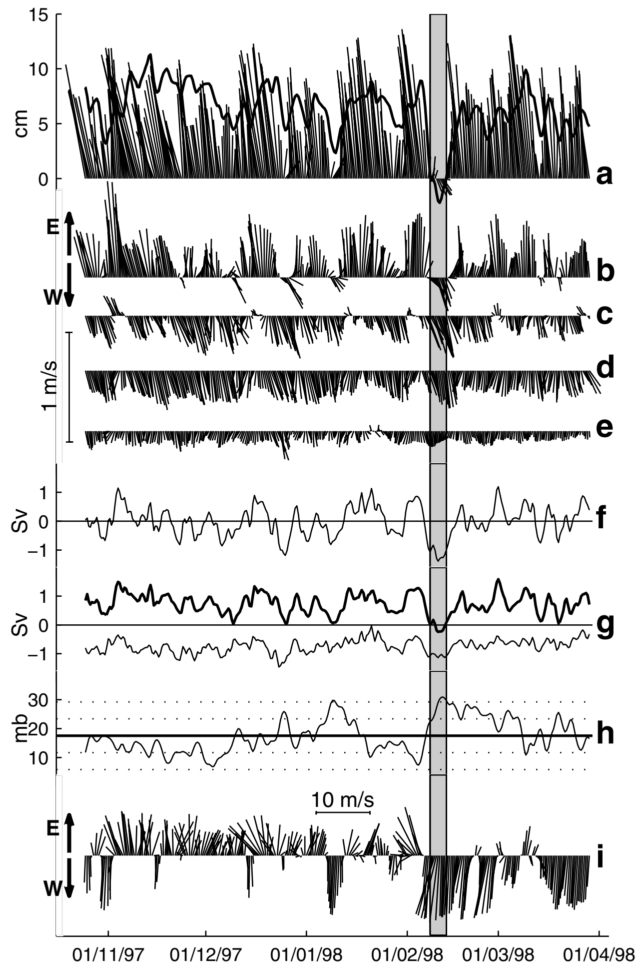

Figure 2. Time series of (a) velocity at depth 1 in the mooring site. Solid line is the across-strait sea level difference (left scale), (b), (c), (d), and (e) velocity at depthes 2, 3, 4, and 5 respectively, (f) net flow, (g) inflow (thick line) and outflow (thin line), (h) spatially atmospheric pressure (mb – 1000) over the Mediterranean Sea. The horizontal thick line is the mean (1017.5 mb) and dotted lines mark the mean ± std and mean ± 2std, std being the standard deviation of 5.8 mb. (i) averaged wind speed in the Strait of Gibraltar.

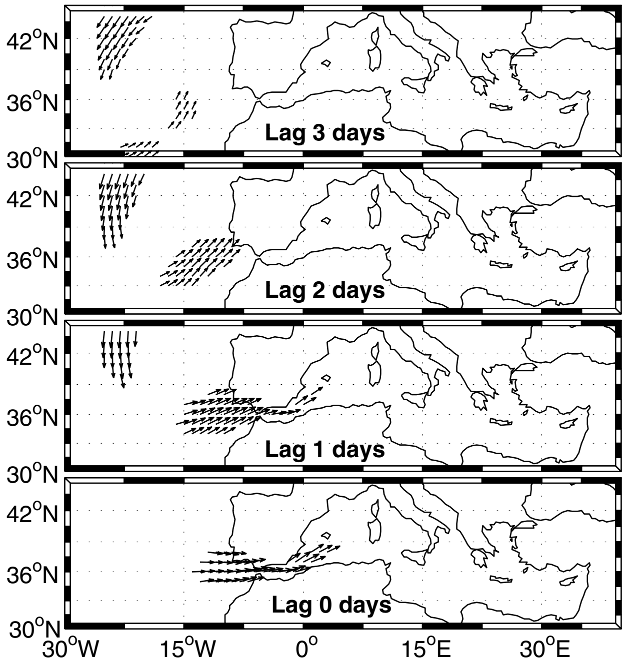

Figure 3. Covariance of the inflow through the Strait of Gibraltar and wind-stress at different lags.

pattern depicted in Figure 3 reflects the typical west-to-east propagation of atmospheric systems but also supplies a quantitative idea of the wind field spatial scale that induces inflow variations and of the most favourable direction for it. Zonal winds over the Gulf of Cádiz and Alboran Sea at 0 lag are the most efficient. This is better seen in Figure 4, which shows contours of the correlation coefficient between the projection over different angles of the spatially-averaged wind stress in the Gulf of Cadiz and Alboran Sea and the inflow through the Strait. Maximum correlation is found at lag 0 for angles around 10 degrees North from East, in agreement with Figure 3.

[12] The easiest way to address the inflow response to wind would be a vertically integrated model in which wind-stress, τ_w, cancels interfacial and lateral/bottom friction, τ_f. In this case, using the usual relationship $\tau = C_D\rho\,|u|u$ for wind and bottom stresses, it is easy to show that $Q_1 = a.U_{wind}$, with $a = (C_{Da}\rho_a/C_{Dw}\rho_w)^{1/2}WH$. Here C_D is the drag coefficient, subindices a and w refer to air and water, respectively, and W and H are the width and depth of the layer. If $C_{Da} \approx C_{Dw}$, ρ_a = 1.3 kg.m^{-3}, ρ_w = 1028 kg m^{-3}, W = 18 km at the eastern section (Figure 1) and $H \approx$ 100 m which corresponds to submaximal rather than maximal exchange [*García Lafuente et al.*, 2002b], then $Q_1 \approx 0.06\ U_w$ Sv if wind speed U_w is given in m s^{-1}. This represents a rather large gain to wind stress.

[13] The main deficiency of this approach is that wind fluctuates at time scales too short (typically 5 days) to reach the steady state implicitly assumed in the formulation given above. Spin up time under the action of wind stress in this vertically integrated model could be estimated as $T_{spin} = H(1/2f\mu)^{-1}$ [*Gill*, 1982], where μ is the vertical eddy-viscosity and f = 8.5 10^{-5} s^{-1} is the Coriolis parameter at 36°N. With μ = 2.5 10^{-3} m^{2}s^{-1} in the upper layer [*Wesson and Gregg*, 1994] and $H \approx$ 100 m, $T_{spin} \approx$ 3.5 days, which is the order of the time-scale of wind fluctuations. Thus the inflow response is frequency dependent. The linearised along-strait momentum equation that includes wind stress and interfacial friction could be written [*Gill*, 1982, p. 414] as

$$\frac{\partial \boldsymbol{u}}{\partial \boldsymbol{t}} - \boldsymbol{f}\boldsymbol{v} = -\boldsymbol{g}\frac{\partial \eta}{\partial \boldsymbol{x}} + \frac{\tau_{wind}}{\rho \boldsymbol{H}} - \sqrt{\frac{\boldsymbol{f}\mu}{2\boldsymbol{H}^2}}\boldsymbol{u} \tag{1}$$

[14] Coriolis acceleration vanishes since v = 0. If the along strait sea level gradient is ignored to analyse the balance between stresses and inertia terms, a time dependence $\exp(-i\omega t)$ for the variables is assumed and, finally, wind stress $\tau = C_{Da}\rho_a|U_w|U_w$ with C_{Da} = 1.5 10^{-3} is linearised taking a representative value for $|U_w|$, say the root mean square wind speed amplitude $U_{0w}/2^{1/2}$, then $Q_1 = G(\omega)U_{0w}^2$, $G(\omega)$ being the complex gain whose modulus, G_m, and phase, G_φ, are given by:

$$\boldsymbol{G_m} = \frac{\boldsymbol{C}_{Da}\rho_a}{\sqrt{2}\rho_{\boldsymbol{w}}} \cdot \frac{\boldsymbol{W}}{\sqrt{\frac{\boldsymbol{f}\mu}{2\boldsymbol{H}^2} + \omega^2}} \tag{2}$$

$$\boldsymbol{G}_\varphi = \tan^{-1}\left(\frac{\sqrt{2}\omega \boldsymbol{H}}{\sqrt{\boldsymbol{f}\mu}}\right) \tag{3}$$

[15] Figure 5 shows contours of $G_mU_{0w}^2$ within the frequency and wind speed intervals of interest, for different values of H. During the event of flow reversal, the wind gust would be characterised by U_{0w} = 10 m s^{-1} and around 10 days of period, which gives a wind induced inflow variation around 0.3 Sv, somewhat lower (0.16 Sv) if frequency is doubled (period of 5 days). The inclusion of the inertial term for subinertial frequencies reduces the inflow response by a factor greater than 2 with regards to the more simple model which only balances wind stress and friction.

5. Discussion and Conclusions

[16] When describing the inflow reversal, it was argued that the strong atmospheric pressure variation over the Mediterranean Sea that triggered the event was not enough

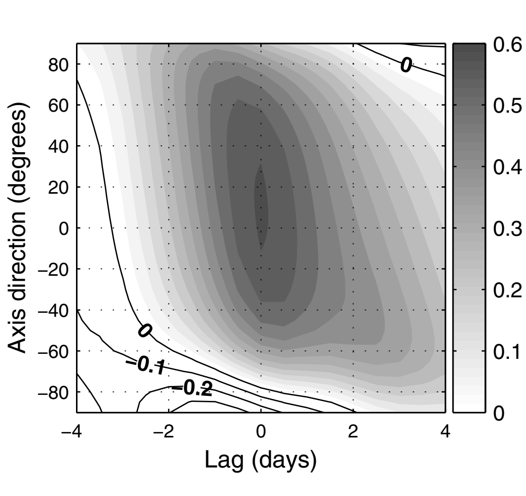

Figure 4. Contours of the correlation coefficient between wind-stress for different angles and inflow through the Strait of Gibraltar. Positive correlation has been plotted as filled contours (right scale).

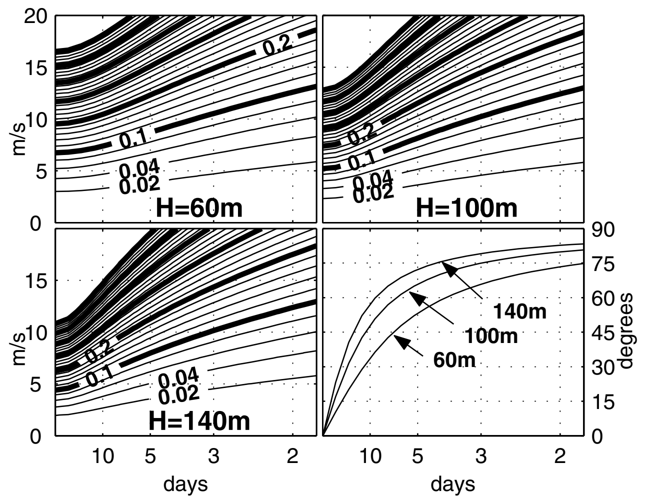

Figure 5. Contours of inflow gain as a function of frequency and wind speed for different thickness of the upper layer. Bottom-right panel is the phase for the 3 thicknesses shown in the first three panels.

to cancel the mean inflow. Estimates based on previous works [*Candela et al.*, 1989; *García Lafuente et al.*, 2002a] predicted an inflow variation of 0.5 to 0.6 Sv, which is only 75% the mean inflow of 0.8 Sv. The simple analysis of wind forcing carried out in the previous section showed that wind stress during the event could have induced an additional fluctuation of up to 0.3 Sv of the same sign as induced by the atmospheric pressure. Both contributions add up to the mean flow and explain the observed reversal.

[17] There are some points worth mentioning. First of all it should be emphasised that atmospheric pressure is the main force for both net flow and inflow fluctuations. This is seen in Figure 2, noticing that once the atmospheric pressure reached its maximum and started to diminish, that is, once the atmospheric pressure force changed sign, the current recorded by the uppermost current meter veered again to the usual eastward direction, despite the fact that easterlies were still blowing over the area. Secondly, the simultaneous occurrence of high pressure over the Mediterranean and easterlies in the Strait of Gibraltar area, circumstances which can interrupt the inflow, is not infrequent but a consequence of the motion of atmospheric systems at these latitudes. Highs moving westwards and entering the Mediterranean Sea usually leave easterlies behind. It is the conjunction of rapid and large atmospheric pressure changes and strong and persistent easterlies that are not often found. A situation that resembles the event discussed here took place during the first days of January 1998 (Figure 2). Neither was the atmospheric pressure change so strong, nor the easterlies so persistent as in February but, yet again, the eastward velocity at 40m depth and the sea level difference were greatly reduced and the inflow nearly reverses. Actually, the current meter at 70m depth registered westward currents, suggesting that the interface was somewhere between 70m and 40m depth. Finally, winter is the season when atmospheric pressure fluctuations are the strongest and therefore the season that propitiates inflow interruptions.

[18] Inflow variations can produce drastic changes in the surface circulation of the adjacent Mediterranean basin. *Bormans and Garrett* [1989] showed that if the inflow velocity falls below a value such that its inertial radius is less than the curvature radius of the south eastern corner of the Strait, then the inflow forms a coastal jet attached to the African coast in the Alboran Sea and stops feeding the Western Alboran anticyclonic Gyre. It has been speculated that the formation of this coastal jet could cause the Western Alboran gyre to disappear, or be the seed for generating a newer gyre that would co-exist with the older one and with the Eastern gyre, giving rise to a three-anticyclonic-gyre situation [*Viudez et al.*, 1998]. Inflow interruptions or inflow weakening like the one just mentioned, which are observed under strong meteorological forcing in winter, are disturbing events that could lead to the formation of the coastal jet mode. Taking into account the characteristic growth time of 30 days for a new gyre to fully develop [*Gleizon et al.*, 1996], the repetition of events, not necessarily so extreme like the one described here, but still strong enough to diminish the inertial radius below that critical value, could maintain the jet in coastal mode for long periods in winter.

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