Seasonal variability of the surface inflow through the Strait of Gibraltar

Seasonal variability
Sea level
Gibraltar
Hydraulic control
Mediterranean Sea

Variabilité saisonnière
Niveau de la mer
Gibraltar
Contrôle hydraulique
Mer Méditerranée

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ABSTRACT

We investigate the seasonal variability of the surface inflow through the Strait of Gibraltar, using historical data of sea level, wind stress, temperature and salinity, as well as idealised models and simple physical arguments.
The seasonal changes, deduced from monthly mean sea level differences across the Strait, do not reflect: a) a purely barotropic flow as required by mass conservation; b) an adjustment, month-by-month, of a two-layer salt-conserving flow; or c) a baroclinic flow that is hydraulically controlled at the sill and driven by density changes of the inflowing Atlantic water without changes in interface depth.
We suggest that the exchange through the strait is submaximal, and that the interface depth changes over the course of the year. We attribute part of this change to the baroclinic set-up and set-down associated with seasonal variations in the wind stress, and the remainder to partial draining of the reservoir of dense outflowing Mediterranean water during the summer when it is not being formed.

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RÉSUMÉ

Variabilité saisonnière du flux entrant dans le détroit de Gibraltar

Nous avons examiné la variabilité saisonnière du flux entrant dans le détroit de Gibraltar, à partir de données du niveau de la mer, de la tension du vent, de la température et de la salinité, interprétées au moyen de simples modèles physiques.
Les variations du flux de surface sont déduites des moyennes mensuelles de la différence du niveau de la mer de part et d'autre du détroit. Ces variations ne peuvent s'expliquer, ni par un écoulement purement barotrope, ni par un écoulement barocline satisfaisant une conservation mensuelle du bilan en sel, ni par un écoulement barocline soumis à un contrôle hydraulique au seuil du détroit et induit par les variations de densité des eaux atlantiques y pénétrant, sans variation saisonnière de la profondeur de l'interface entre les deux couches.
Nous sommes donc amenés à suggérer que l'échange à travers le détroit est inférieur à l'échange maximum, et que la profondeur de l'interface varie au cours de l'année. Nous attribuons cela, d'une part à la dénivellation barocline due aux variations saisonnières de la tension du vent, d'autre part à un drainage partiel des eaux méditerranéennes durant l'été lorsque celles-ci ne sont pas formées.

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INTRODUCTION

The flow through the Strait of Gibraltar is one of the classic problems of oceanography (*see* Defant, 1961; and Deacon, 1971). The simplest view of the flow is that it is a two-layer "inverse estuarine" circulation, with North Atlantic water flowing in at the surface and saltier Mediterranean water flowing out at depth, in response to an excess of evaporation minus precipitation over the Mediterranean Basin. Given the latter, and the salinity difference between the two layers, mass and salt conservation then imply an inflow of about 1.7×10^6 m^3 s^{-1} and slightly less outflow (*e.g.* Bethoux, 1979). This is reasonably compatible with the measured transports reported by Lacombe and Richez (1982) of about 1.2×10^6 m^3 s^{-1}. However, these estimates of

transport are not accurate enough to combine with conservation laws and so provide a check on the rather uncertain values of evaporation minus precipitation (Bunker *et al.*, 1982).

It is only recently that an attempt has been made to explain the observed salinity difference between the two counterflowing layers in the strait. Bryden and Stommel (1984) argue that strong mixing in the Mediterranean drives the system to an "overmixing" limit in which the exchange is maximized subject to hydraulic control at the sill in the strait. This solution minimizes the salinity difference between the two layers and gives a mid-depth interface above the sill.

Farmer and Armi (1986) also maximize the exchange for flow that is critical at the sill, but add the further condition that the flow be hydraulically connected to conditions away from the sill, in contrast to the Bryden and Stommel solution. They argue that there is a second control point at the contraction near the exit of the surface flow into the Mediterranean. The exchange is maximum only when the flow is also critical at the contraction, in which case the interface over the sill is at its highest level but still lower than mid-depth. There is thus somewhat less exchange than that predicted by Bryden and Stommel (1984); their solution, with a mid-depth interface, would give more exchange but is physically unrealisable.

Farmer and Armi (1986) also consider the effect on the exchange of the significant barotropic fluctuations (such as tides) in flow through the strait and it is clear that these will have to be taken into account in any complete model. For the moment, though, the most uncertain qualitative assumptions of these theories are those involving maximization of the exchange, either via the "overmixing" argument of Bryden and Stommel (1984), or the discussion of other control points by Farmer and Armi (1986).

In this paper we shall discuss seasonal changes in the surface inflow through the strait, motivated by historical sea level data which show small but significant changes ($\simeq 6\%$ of the mean) over the course of a year. We consider it important to study these changes in the flow as this may help to confirm or disprove the hypotheses being used in models for the mean flow and so improve our ability to predict flow variations due to changing conditions in the Atlantic or Mediterranean. Also, in the presence of seasonal and interannual variability in the flow, matching part of a long record of a readily observable variable (*e.g.* coastal sea level) to results from a detailed one-or two-year observational program may help us define more accurately the long term mean and variability of the flow.

We shall show later that the seasonal variations in surface inflow, established in the next section, do not reflect: a) a purely barotropic flow; b) a baroclinic salt-conserving flow; or c) a baroclinic flow that is hydraulically controlled at the sill and driven by density changes of the inflowing Atlantic water without changes in interface depth.

This suggests that, within the context of the two-layer model used, the interface depth must change over the course of the year. This change in interface depth may be partly due to the effect of wind, as discussed later, or, more interestingly, to the draining of the outflowing Mediterranean water during the season when it is not being formed.

The paper concludes with a discussion of the implications of our interpretation and suggestions for further work that would clarify the nature and cause of the seasonal variations and, more importantly, the mean flow.

EVIDENCE FOR SEASONAL VARIATION OF THE INFLOW

Our indirect evidence for a seasonally changing surface inflow is based on sea level observations from both the North Atlantic and Western Mediterranean. We have analyzed all available monthly mean sea level data for the period 1950-75 from the 9 tide gauge positions in Figure 1 (all data were obtained from the publications of the Permanent Service for Mean Sea Level; the data periods are shown in Fig. 2). The influence of local air

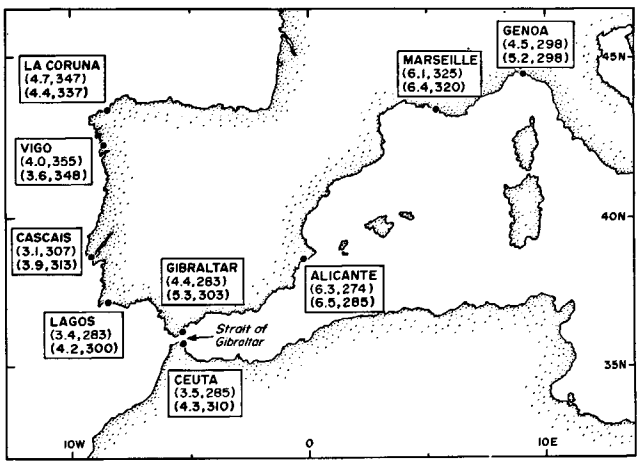

Figure 1

Amplitude and phase of the annual cycle of mean sea level (upper line) and of sea level + air pressure (lower line) at 9 locations. The data periods are shown in Figure 2. Zero phase corresponds to 1 January.

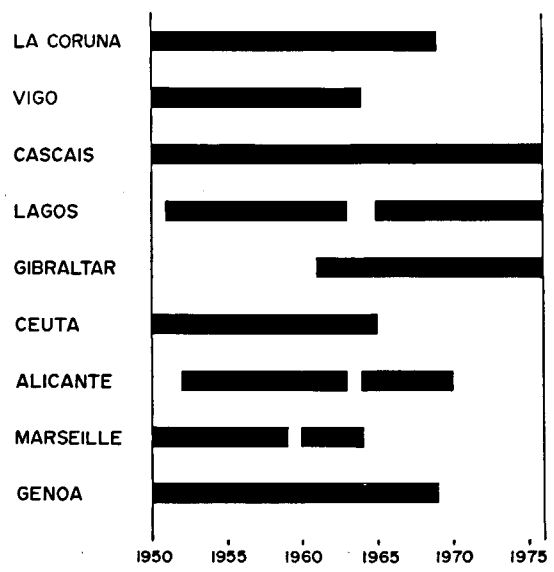

Figure 2

Monthly mean sea level data periods.

Table 1
Amplitude and phase of the annual and semi-annual cycles for different quantities. Zero phase corresponds to 1 January.

	Sa		*Ssa*	
	Amplitude	*Phase Lag*	*Amplitude*	*Phase Lag*
$\Delta_x\eta'$ (cm)	2.5	260°	0.4	260°
$\Delta_y\eta'$ (cm)	1.1	275°	0.4	73°
(E-P)′(cm/month)	7.1	181°	2.5	46°
ρ_1' (kg/m³)	0.31	59°	0.01	76°
ρ_2' (kg/m³)	0.02	65°	0.04	100°
$\rho_2'-\rho_1'$ (kg/m³)	0.29	238°	0.03	289°

pressure has been removed from each monthly record by simply adding sea level (cm) and the local air pressure (mb). Although the air pressure correction is small ($\simeq 1$ cm) it reduces the scatter in the amplitudes and phases of the annual cycle; in the North Atlantic, for example, the maximum differences in amplitude and phase between tide gauges are reduced from 1.6 cm and 72° to 0.8 cm and 48° respectively (in the rest of this paper we shall denote the monthly mean total pressure (sea level + air pressure) simply by η for convenience. We shall use the term "seasonal variation" to describe the average, over many years, of the month-to-month changes, and the term "annual cycle" to describe the Fourier component of the seasonal variation with a period of exactly one year).

The amplitude of the annual cycle of η along the coast of Portugal is about 4 cm; this is considerably less than the 6 cm amplitude found in the Western Mediterranean (Fig. 1). There is also a significant difference in amplitude across the strait (Fig. 1, Tab. 1). The purpose of this section is to infer the seasonal variation in the surface inflow through the strait from these seasonal changes in sea level.

Cross-strait variations

The difference in the seasonal variation of η across the strait ($\Delta_y\eta' = \eta'_{\text{Gibraltar}} - \eta'_{\text{Ceuta}}$) is shown in Figure 3. Primed quantities denote the seasonal variation. The seasonal variations at both Gibraltar and Ceuta are based on different averaging periods (Fig. 2). In fact the common period was only 36 months and so it was not possible to estimate the standard error of each of the 12 monthly averages of $\Delta_y\eta'$ directly. However, we have estimated the error indirectly by first calculating the differences between the two time series over their common period, subtracted the mean seasonal variation in the difference (Fig. 3) and then determined the standard deviation (σ) of the 36 residual values. The standard error of the monthly averages in Figure 3 was then approximated by $\sigma/(n)^{1/2}$ where $n(\simeq 14)$ is the number of years used to form each seasonal variation curve separately.

Both the well-defined seasonal variation in $\Delta_y\eta'$ and the relatively small standard error (Fig. 3) suggest that there is a real seasonal variation in the pressure difference across the strait. In order to interpret this pressure difference in terms of inflow variations we integrate the geostrophic relationship to obtain

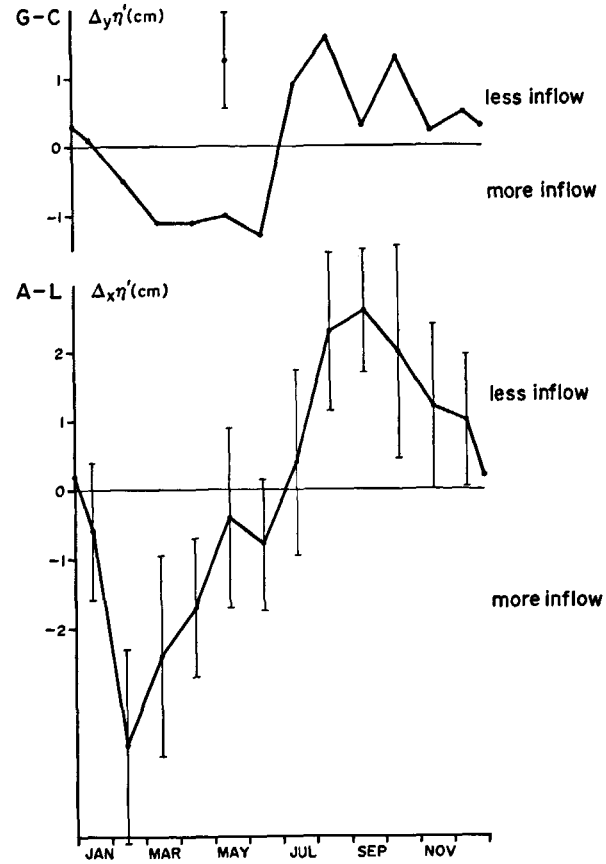

Figure 3
Seasonal variation in the difference between Gibraltar and Ceuta (above) and between Alicante and Lagos (below), together with standard errors.

$$u_1'/\bar{u}_1 = -g\,\Delta_y\eta'/f\,W\bar{u}_1 = \Delta_y\eta'/\Delta_y\bar{\eta} \qquad (1)$$

where u_1 denotes the eastward velocity of the top layer and W the effective width at the GC section (Fig. 4), and overbar and prime denote mean and seasonally fluctuating quantities respectively. The Coriolis parameter f is taken to be $0.85 \times 10^{-4}\ s^{-1}$.

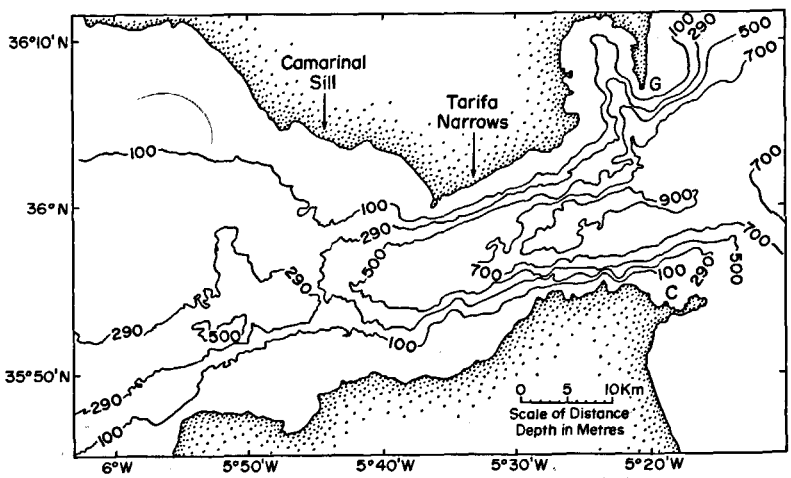

Figure 4
Bathymetry of the Strait of Gibraltar (Instituto Español de Oceanografia, 1983). The locations of the tide gauges at Gibraltar and Ceuta are marked by G and C respectively.

From Table 1 the amplitude of the annual cycle in $\Delta_y\eta'$ is 1.1 cm with a maximum in September. This corresponds to an annual cycle in geostrophic inflow with an amplitude of 7 cm s^{-1}. The tide gauges at Gibraltar and Ceuta have not been levelled with respect to each other and so we do not know the mean gradient in total pressure $\Delta_y\bar{\eta}$. Current meter data from one station between Gibraltar and Ceuta however indicate a mean surface inflow $\bar{u}_1$ of 1.2 ms^{-1} (Lacombe, Richez, 1982), though from only 2 days of data. Together with W = 18 km, this implies, from (1), an amplitude of 6% (to within, perhaps, a factor of 2) for the annual cycle of $u_1'/\bar{u}_1$ at the GC section, with strongest inflow in the spring.

Along-strait variations

The seasonal variation along the strait ($\Delta_x\eta' = \eta'_{Alicante} - \eta'_{Lagos}$) is shown in Figure 3. Data came from the common period 1952-62,1965-69 and so we have been able to calculate the standard error of each of the 12 monthly averages. There is clearly a statistically significant annual cycle in $\Delta_x\eta'$. From Table 1, the semi-annual cycles of both $\Delta_x\eta'$ and $\Delta_y\eta'$ are much smaller than the respective annual cycles. The difference in mean sea level between Alicante and Lagos ($\Delta_x\bar{\eta}$) has been determined from geodetic levelling to be 19 cm (Levallois, Maillard, 1970), and the difference in mean air pressure between Alicante and Lagos is small ($\simeq$1 mb). Taken together, these observations imply a mean drop in total head from Lagos to Alicante of about 20 cm. Substituting 2.5 cm for the amplitude of the annual cycle of $\Delta_x\eta'$ (Tab. 1) we find a 12% amplitude for the annual variation of $u_1'/\bar{u}_1$ if we assume that the surface inflow through the strait is proportional to the head between Lagos and Alicante. This is not much different from the (rather uncertain) 6% variability deduced from cross-strait differences. Moreover, the phase lags (Tab. 1) are in good agreement, both indicating maximum inflow near the end of March.

This similarity is encouraging, but, as we shall see later, removal of the wind effect from both the Gibraltar-Ceuta and Alicante-Lagos pressure differences leaves residual cycles that are out of phase. This suggests that variations in the sea level difference between Alicante and Lagos may not be a good measure of variations in the sea level difference between each end of the strait (although we will not pursue it here, it is worth remarking that the annual cycles of residual sea level at Lagos and Alicante, after removal of the effect of wind, are very similar in amplitude and phase to the annual cycles of dynamic height, in the nearby Atlantic and Mediterranean respectively, computed by Pattullo *et al.* (1955). Hence there may be seasonal changes in features, such as the Alboran gyre, near but outside the strait).

In spite of these reservations about the value of the Alicante and Lagos data, the Gibraltar and Ceuta data alone provide fairly convincing evidence for a seasonal variation, with an annual cycle of amplitude roughly 6% and maximum inflow in the spring. For the remainder of the paper, we will use the observed fluctuations in the sea level difference between Gibraltar and Ceuta to infer, from the geostrophic formula, the seasonal variations of the surface inflow.

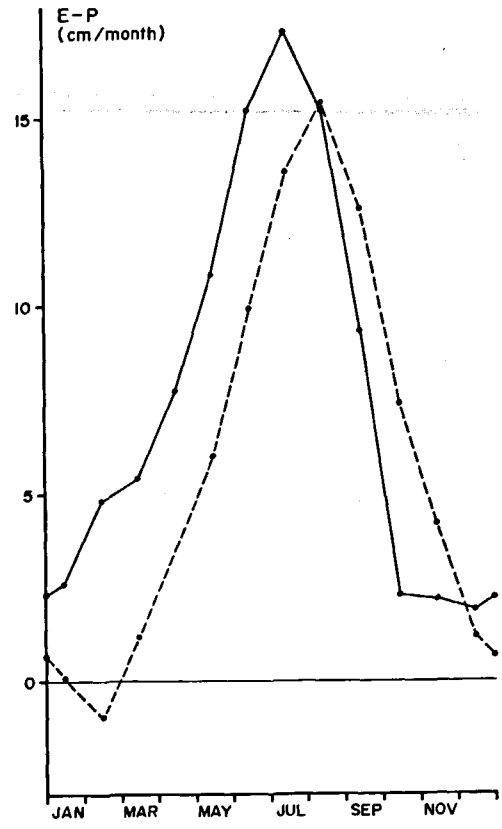

Figure 5
Seasonal variation in evaporation-precipitation over the Mediterranean from the World Survey of Climatology, 1970 (*solid line*) *and from Carter,* 1956 (*dashed line*).

SIMPLE MODELS

The simplest model for the flow through the Strait is a two-layer exchange with relatively light Atlantic water flowing into the Mediterranean above an outflow of denser Mediterranean water (the subscripts 1 and 2 refer to the upper and lower layer respectively, ρ being density, *u* the eastward current, *h* the thickness of each layer and W the width of the strait). An excess of evaporation (E) over precipitation (P) over the Mediterranean (of surface area A) requires a net inflow of Atlantic water to achieve the mass balance

$$W(\rho_1 u_1 h_1 + \rho_2 u_2 h_2) = \iint_A [\rho_0(E-P) + \rho_3\, \partial/\partial t\,(\Delta\zeta)]\, dA \qquad (2)$$

where $\Delta\zeta$ is the difference between the observed sea level and the steric height (with respect to a depth of no change with density ρ_3) and ρ_0 is the density of fresh water. We have ignored the negligible contributions of the inflows from rivers and the Black Sea (Carter, 1956; Tixeront, 1970).

There are few reliable estimates of the seasonal variation of E-P (Carter, 1956; Bunker, 1972); even longterm means, averaged over the whole Mediterranean, vary from 5 cm/month (Carter, 1956) to

12 cm/month (Sverdrup *et al.*, 1942). The E-P data used in this study were obtained from the World Survey of Climatology (Wallen, 1970). We simply averaged the monthly estimates of E-P for all available stations and scaled the results to match the mean of 8 cm/month obtained by Bethoux (1979). The resulting monthly mean E-P cycle is shown in Figure 5 along with Carter's estimates. The amplitude of the annual cycle in E-P is 7 cm/month, with a much smaller semi-annual component (Tab. 1).

The second term on the RHS of (2) is simply the rate of change of the difference between sea level variations and steric height. Pattullo *et al.* (1955) provide estimates of the annual cycle in both these quantities and from them we can infer that this term has an annual amplitude generally less than 2 cm/month, and is thus small in comparison to E-P for the annual cycle. We ignore it here; including it would not affect our conclusions. Finally, if we also ignore small density differences, the equation of mass conservation may be approximated by

$$u_1 h_1 + u_2 h_2 = 1/\mathrm{W} \iint_A (\mathrm{E}-\mathrm{P})\, dA. \tag{3}$$

Barotropic response

It is straightforward at this point to show that the seasonal fluctuations in the inflow are not due to a barotropic change in the flow. We denote the barotropic response by

$$u_b = 1/\mathrm{WH} \iint_A (\mathrm{E}-\mathrm{P})\, dA \tag{4}$$

where $\mathrm{H} = h_1 + h_2$ is the total depth. At the GC section, using $\bar{u}_1 = 1.2\ \mathrm{ms}^{-1}$ and the appropriate W and H, we obtain $u_b'/\bar{u}_1 = 0.4\%$ (Tab. 2). This is clearly much less than the value of about 6% for $u_1'/\bar{u}_1$ obtained from the sea level data and implies that the seasonal response of the inflow to the E-P fluctuations must be baroclinic.

Table 2

Amplitude and phase of the annual cycle of $u_1'/\bar{u}_1$. Zero phase corresponds to 1 January.

	Amplitude	*Phase Lag*
From observations		
Using G-C	6%	95°
Using A-L	12%	80°
From theory		
$u_b'/\bar{u}_1$	0.4%	181°
$u_s'/\bar{u}_1$	80%	181°
$u_\rho'/\bar{u}_\rho$	6%	236°

Salt conservation

It is also straightforward to show that a baroclinic response which satisfies mass conservation (3) and salt conservation, *i.e.*

$$u_1 h_1 \mathrm{S}_1 + u_2 h_2 \mathrm{S}_2 = 0 \tag{5}$$

from one month to the next is also inappropriate. Combining (3) with (5), we obtain the velocity of the upper layer from the familiar Knudsen formula

$$u_s = \mathrm{S}_2(\mathrm{S}_2 - \mathrm{S}_1)^{-1}\, \mathrm{H}\, h_1^{-1}\, u_b \tag{6}$$

with u_b defined by (4). Observations show that at the GC section, $\mathrm{S}_2/(\mathrm{S}_2\text{-}\mathrm{S}_1) = 20$ and $\mathrm{H}/h_1 = 10$. Hence, $u_s = 200\, u_b$ and $u_s'/\bar{u}_1 = 80\%$ (Tab. 2), again clearly inconsistent with the $\pm 6\%$ fluctuation in $u_1'/\bar{u}_1$ inferred from sea level differences. This implies that the baroclinically varying flow does not conserve the salt content of the Mediterranean on a monthly basis, a conclusion which is unaffected by plausible seasonal changes in h_1 and S_1.

Hydraulic considerations

Recent models for the inflow involve the dynamics of two layer exchange through the strait. In particular, Bryden and Stommel (1984) have analyzed the mean exchange in terms of the following condition of hydraulic control at a sill:

$$u_1^2/h_1 + u_2^2/h_2 = 4\, u_\rho^2/\mathrm{H}_0 \tag{7}$$

where H_0 is the total depth $h_1 + h_2$ at the sill and u_ρ, defined by

$$u_\rho = [g\, \rho_2^{-1} (\rho_2 - \rho_1)\, \mathrm{H}_0/4]^{1/2} \tag{8}$$

is the speed at which each layer would advance in the lock exchange problem (Turner, 1973, p. 70). Bryden and Stommel (1984) use the concept of "overmixing" and (7) to argue that extensive mixing in the Mediterranean drives the density difference (ρ_2-ρ_1) to a minimum and, consequently, the exchange to a maximum. They found that the interface at the sill resides at mid-depth (*i.e.* $h_1 = 0.50\ \mathrm{H}_0$ or 142 m) for maximal exchange.

Farmer and Armi (1986) have also examined the two layer exchange through a strait that is critical at a sill. However, in contrast to Bryden and Stommel, they hydraulically connected the flow through the strait to an adjacent contraction using simple continuity and Bernoulli relations. In the context of the Strait of Gibraltar, the sill and contraction would correspond to Camarinal Sill and Tarifa Narrows respectively (Fig. 4). The maximal exchange in the Farmer and Armi model occurs when the flow at the contraction also becomes critical. In this state of maximal exchange the interface is lower than mid-depth over the sill. In fact, using the observed contraction ratio of 0.8 (width at Tarifa Narrows/width at Camarinal Sill) the Farmer and Armi model implies $h_1 = 0.67\ \mathrm{H}_0$ (or 190 m) at Camarinal Sill and $h_1 = 0.37\ \mathrm{H}_0$ (or 105 m) at Tarifa Narrows, and gives a maximal exchange which is 28% less than that given by the Bryden and Stommel (1984) theory.

In the rest of this paper we will use the general theory of Farmer and Armi to model the seasonal variability of the inflow. However before applying the theory it is important to note that it allows at least two different types of solution. The first type occurs when the flow at the contraction, like that at the sill, is critical. In this state of maximal exchange, the flow is supercritical to the east of the contraction and has an increasingly shallow and fast-moving upper layer as the strait deepens and widens. A sketch of the interface along

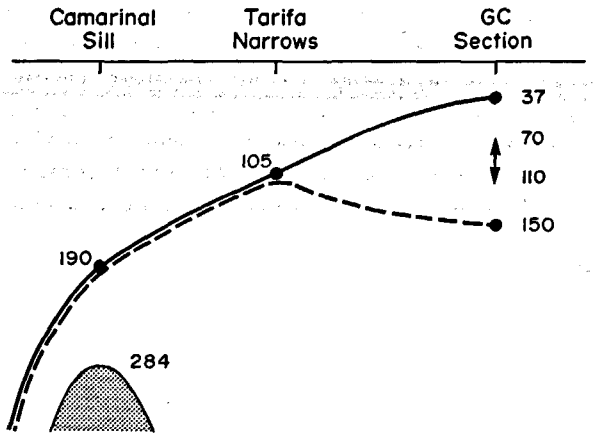

Figure 6
Sketch of the interface position for maximal exchange (supercritical flow at the GC section, solid line) and for slightly submaximal exchange (subcritical flow at the GC section, dashed line). The observed range, from 70 to 110 m is also indicated.

the strait is shown in Figure 6. The supercritical region essentially isolates conditions at the contraction and at the sill from changes in interface depth in the Mediterranean. Using the observed width increase from Camarinal Sill to the GC section (50%) and observed g' (2×10^{-2} ms^{-2}), the maximal exchange solution predicts $h_1=37$ m, $u_1=2.2$ ms^{-1} and a squared surface layer Froude number of 6.5 at the GC section. It is possible that such a supercritical inflow could join the sluggish Mediterranean circulation through a hydraulic jump somewhere between the contraction and the GC section. However, the hydrographic data described below do not support this idea and it is not considered further.

A second type of solution from the Farmer and Armi model occurs when the flow at the contraction is subcritical (though it is still critical at the sill) and the exchange is submaximal (this condition is sometimes described as the exit control being flooded; it occurs if the interface in the Mediterranean is deeper than $0.56\ H_0=159$ m for the parameters described above). The flow remains subcritical to the east of the contraction with a slowing and deepening upper layer. Assuming the exchange to be slightly submaximal, then $h_1=150$ m, $u_1=0.54$ ms^{-1} and the squared surface layer Froude number is 0.1 at the GC section (Fig. 6). For submaximal exchange it is possible for changes in interface depth in the Mediterranean to alter conditions at the sill.

The observational data base is somewhat limited and it is difficult to determine if the exchange is indeed maximal *i.e.* it is difficult to determine which of the two flows depicted in Figure 6, is appropriate. For example, measurements of the depth of the upper layer at the GC section in mid-strait, described by Cavanie (1973), give a 4-day mean of about 110 m. More recent reports by Lacombe and Richez (1982; 1984) show that the mean interface depth is approximately 70 m in mid-strait and varies from 50-90 m over a tidal cycle. Clearly the above depths of the upper layer at the GC section do not correspond to either of the possibilities shown in Figure 6 (*i.e.* slightly submaximal or maximal exchange).

We believe that this difficulty in reconciling observations and theory can be resolved by considering the effect of interfacial friction. More specifically, we believe that friction could be responsible for raising the interface by about 50 m from the slightly submaximal branch at the GC section into the observed range. Some support for our hypothesis comes from the mean sea level data reported by Levallois and Maillard (1970); they found a drop in mean sea level of 15 cm from Cadiz to Malaga, much more than the 5 cm attributable to the Bernoulli effect. The remaining 10 cm drop in sea level along the strait is consistent with a 50 m rise in interface height if the pressure gradient in the sluggish lower layer is assumed small. It is perhaps appropriate to note at this point that this balance between the sea level gradient and interfacial friction, over 35 km, requires a quadratic drag coefficient of about 3×10^{-3} for $u_1=1$ ms^{-1}, $h_1=100$ m. This is an order of magnitude larger than the values normally quoted (Csanady, 1978*a*), but it may not be unrealistic in a region with large internal tides and other high frequency motions.

The above interpretation of existing data is hampered not only by its limited quality and quantity, but also by the idealised model upon which the interpretation is based. In particular, the pronounced vertical shear of the inflowing current at the GC section (Lacombe, Richez, 1982) makes the use of a two-layer model somewhat dubious. Further, it is clear that the effects of the earth's rotation, internal friction and barotropic fluctuations will have to be modelled in a more complete future analysis. However, on the basis of the above discussion, we will assume that the flow at GC is subcritical and use the submaximal exchange solution from the Farmer and Armi theory. The relationship

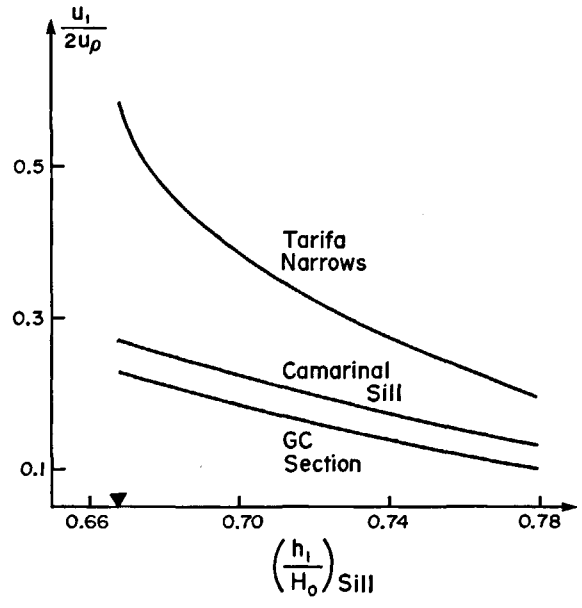

Figure 7
Relationship between the interface depth over the Sill (h_1), the density difference between the two layers (and hence u_p) and the speed of the inflow (u_1) at Tarifa Narrows, Camarinal Sill and at the GC section. The triangle represents the maximal exchange solution at which the interface over the sill is at its shallowest (i.e. $h_1/H_0=0.67$).

between interface depth over the sill, the density difference between the two layers and the speed of the inflow is shown for different sections along the Strait in Figure 7. The shallowest interface depth over the sill is achieved when the flow at the contraction just becomes critical; it is then 0.67 $H_0 = 190$ m deep. Below this minimum depth, the upper layer flow (u_1), the density difference ρ_2-ρ_1 (hence u_ρ) and the interface depth over the sill (h_1) change together as shown in Figure 7. Thus using the submaximal exchange solution of Farmer and Armi we can associate changes in the monthly mean flow past GC (u_1) with changes in u_ρ and/or the interface depth at the sill (and hence in the depth of the interface in the Western Mediterranean). What contributions do u_ρ and h_1 make to the observed annual variation of u_1 inferred from the Gibraltar-Ceuta sea level difference? The dependence of u_1 on u_ρ is linear (Fig. 7) and is considered first.

the annual cycle of $u'_\rho/\bar{u}_\rho$ is of similar magnitude (6%) to the observed inflow variation from sea level data (Tab. 2). However it is clear from Figure 9 that the phase of u'_ρ does not agree with that of u'_1 inferred from sea level data; the latter suggests maximum inflow in about March whereas u'_ρ is maximum in September, the time of minimum upper layer density and maximum ρ_2-ρ_1.

Changes in interface height

Unfortunately there are insufficient hydrographic data from the strait to reliably define the seasonal variation of h_1 over the sill and determine its effect on u_1. Thus our approach has been to combine the sea level and density data (*i.e.* u_1 and u_ρ in Fig. 7) to infer h_1. Using Figure 7, we can estimate h_1 at the sill from u_1 and u_ρ. The latter is known and shown in Figure 9;

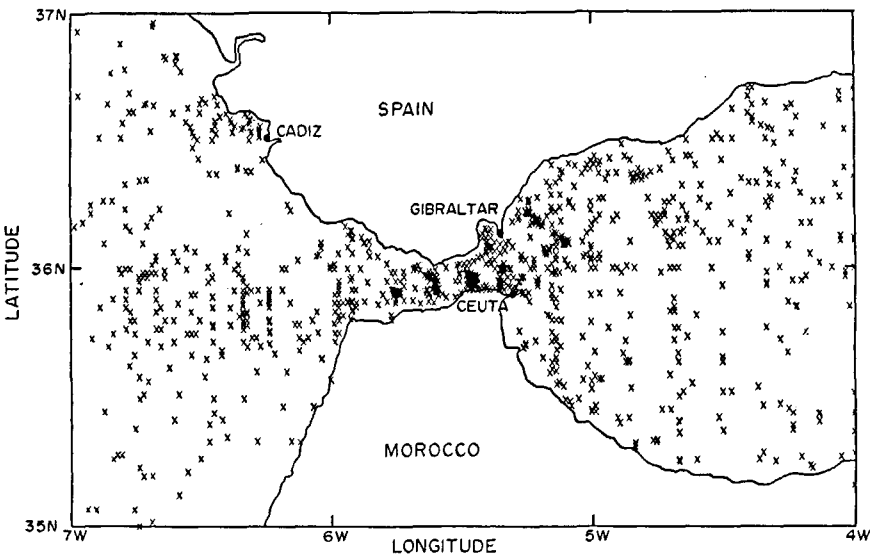

Figure 8
Locations of all hydrographic stations, from 1904 to 1980, used to compute the density.

Density variations

We have calculated the seasonal variation of u_ρ from the observed density difference (ρ_2-ρ_1) between the two layers, using the top 100 m in the Atlantic (averaged over 6° to 7°W) for ρ_1 and 200-300 m in the Mediterranean (averaged over 5° to 5°20′W) for ρ_2. The density data were computed from salinity and temperature data for most of the stations located on Figure 8. They represent all the profiles recorded from 1904 to 1980 available from the National Oceanographic Data Center (NODC). We did not include profiles showing local instability or those taken from depths less than 100 m. We did not find any regional variations in the seasonal cycle of ρ_2-ρ_1 (*e.g.* west Moroccan coast vs. south Spanish coast) and so averaged all the stations in Figure 8 (subject to the above restrictions) to obtain one seasonal variation. The amplitude and phase of the annual and semi-annual cycles of ρ'_1, ρ'_2 and $\rho'_2 - \rho'_1$ are given in Table 1. We note that the seasonal variation in the density difference is almost entirely due to changes in the inflowing Atlantic water.

The seasonal variation in u_ρ, based on the above density data, is shown in Figure 9. The amplitude of

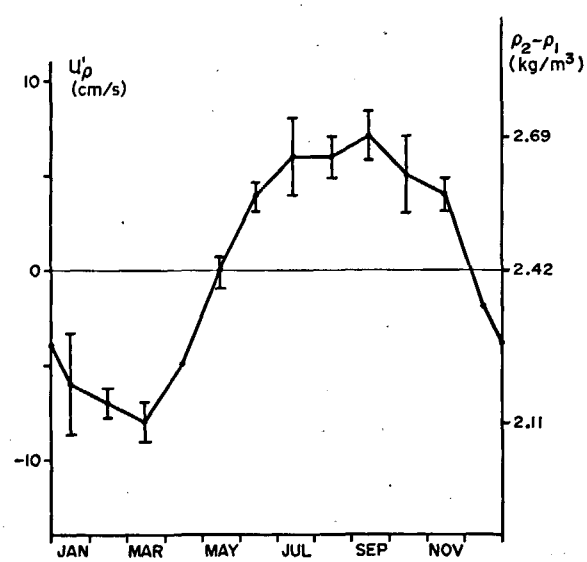

Figure 9
Seasonal variation in the scaling speed u_ρ in response to variations in the observed density difference between the two layers. The standard errors plotted are those due to ρ_1 only. A lack of data for certain months prevented us from including the standard errors due to ρ_2. The contribution of the latter would increase the plotted standard errors by 40% (at most).

the value of $\bar{u}_1$, and hence the absolute values of u_1, are not (the Gibraltar-Ceuta sea level difference only provides fluctuations about the mean). However if we assume $\bar{u}_1 \gg u_1'$ then the seasonal variation of h_1 has a minimum which always occurs in March. Thus fixing u_1 to the maximum exchange value in March determines $\bar{u}_1$ and hence the seasonal variation of h_1 (Fig. 10). We then find that the interface depth over the sill decreases by 21 m from winter to summer. This result is in qualitative agreement with Schott's sparse observations (1928, and reported by Defant, 1961) of seasonal changes in the interface height, although Schott claimed a change of about 80 m. Such changes (Fig. 10) may be evident in the long data sets being obtained in the strait in the 1985-1986 Gibraltar Experiment (Bryden, Kinder, 1985).

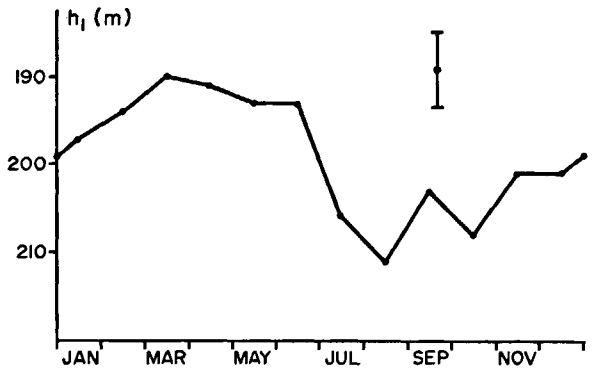

Figure 10
Seasonal variation in the interface depth over Camarinal Sill inferred, via the hydraulic control theory, from the observed sea level difference between Gibraltar and Ceuta and from density data. The standard error is mainly due to $\Delta_y\eta$.

In summary, we hypothesize that the exchange is submaximal and that the seasonal variation of the surface inflow is associated with density and interface depth variations at the sill. The 21 m change of interface depth over the sill corresponds, via the submaximal exchange solution, to an approximate 22 m lowering of the interface in the Mediterranean below its shallowest depth which, in turn, has to be below the flooding value of $0.56\ H_0 = 159$ m. The reasons for the interface variations in the Mediterranean and hence at Camarinal Sill are now examined.

THE EFFECT OF WIND

There are definite seasonal variations in the wind stress, in the neighbouring Eastern Atlantic and Mediterranean Sea (Fig. 11), which might modify the surface inflow in some way.

It is readily shown that any barotropic response to the annual cycle of wind stress (or atmospheric pressure) is, like the barotropic response to E-P discussed earlier, far too small to account for observed changes. However, it is possible that the wind influences the baroclinic exchange through the strait (and hence u_1) by modifying: 1) the density of the inflowing Atlantic water (and hence u_ρ); or 2) the interface depth in the Mediterranean Basin (and hence h_1) as discussed in the previous section.

Before pursuing these ideas with simple quantitative models, we look for statistical evidence for the influence of large scale winds on the total head (adjusted sea level differences) between Gibraltar and Ceuta or Alicante and Lagos. To do this we use a multiple regression model in which the adjusted monthly mean sea level is expressed as

$$\eta = a\tau_x + b\tau_y + c\cos(\omega t - \varphi_1) + d\cos(2\omega t - \varphi_2) + \varepsilon \quad (9)$$

where τ_x, τ_y are the eastward and northward components of wind stress respectively, $\omega = 2\pi\ yr^{-1}$, and ε is a residual uncorrelated with the other inputs. It is important to allow explicitly for an annual cycle (and a semi-annual cycle, though this is much smaller) in this regression; failing to do so in the presence of annual cycles in both adjusted sea level and wind stress

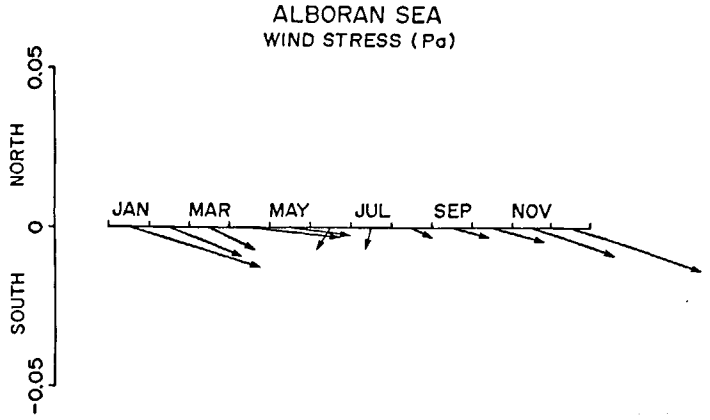

Figure 11
Seasonal variation in the wind stress over the Alboran Sea in the extreme Western Mediterranean (May, 1982).

could lead to a false impression of a causal connection. The model in (9) allows the coefficients a and b to be determined by the aperiodic variability and will give a correct impression of the amount of the annual (and semi-annual) cycle in sea level that can be attributed to the wind stress.

The monthly mean wind stress components are determined from monthly mean pressure data using the method of Thompson and Hazen (1983). This approach allows for the variance in the wind in applying a quadratic drag law, as well as the reduction and backing from geostrophic to surface winds.

The wind gains [(*a,b*) from (9)] for both inside and outside the Mediterranean are shown in Figure 12. The most effective wind direction for raising sea level along the coast of Portugal is alongshore. This direction, and the amplitude, of the wind gain vector appear to be consistent with the arrested topographic wave theory of Csanady (1978 *b*). The most effective wind directions in the Mediterranean are still generally northward but correspond more closely to cross-shore winds. The gains are too large to be explained by wind set-up.

To determine the role of wind forcing in maintaining a pressure head between the Mediterranean and Atlantic, we have also regressed the sea level difference between Gibraltar and Ceuta on the wind in the strait (the overlap period of the Gibraltar and Ceuta sea level records, and hence the data period for this regression analysis, was only 36 months). The resulting gain (Fig. 12) shows that an easterly wind is significantly correlated with an increase in $\Delta_y\eta'$, *i.e.* with decreased inflow (this gain was almost identical to that obtained by subtracting the Gibraltar and Ceuta gains (Fig. 12) derived from separate data periods. In the rest of the paper, we have used the difference in gains as a measure of the wind effect on $\Delta_y\eta'$). A similar regression of the sea level difference between Alicante and Lagos on the wind estimated midway between the two gauges shows that an easterly wind coincides with an increase in Lagos sea level over Alicante.

In Table 3 we show the amplitude and phase of the annual cycles in $\Delta_y\eta'$, $\Delta_x\eta'$, the contributions of the wind to these cycles and the residual cycles. We note that the wind is certainly not responsible for the annual cycle in $\Delta_y\eta'$. From May to October the reduced westerly wind stress (Fig. 11) reduces the inflow and increases $\Delta_y\eta'$ (Tab. 3). The residual annual cycle in inflow inferred from $\Delta_y\eta'$ is of about the same magnitude as the original annual cycle, but peaks in June rather than April.

The $\Delta_x\eta'$ analysis shows a different picture. Indeed, the wind stress fluctuations from the east decrease $\Delta_x\eta'$. This discrepancy in the wind gain between $\Delta_y\eta'$ and $\Delta_x\eta'$ can be explained by a set-up between Alicante and the strait and Lagos which decreases $\Delta_x\eta'$. Again the residual cycle is not much different in magnitude from the original, but its phase is shifted in the opposite direction from $\Delta_y\eta'$. This suggests that the $\Delta_x\eta'$ residual cycle is also affected by dynamical phenomena occurring away from the strait. Alicante and Lagos are probably too far apart to be associated with each end of the strait; stations closer to each end of the strait are needed to represent the along-strait sea level difference. Thus, in accord with our preceeding remarks, we shall concentrate on $\Delta_y\eta'$ alone in our interpretation.

We now determine whether the wind gain for $\Delta_y\eta'$ can be attributed to wind-induced changes in u_ρ or in the interface depth in the Mediterranean. Certainly the direction of the gain vector is as expected; a wind from the east might cause upwelling in the Gulf of Cadiz, hence increasing the density of the Atlantic inflow and so reducing u_ρ and u_1. Alternatively, an easterly wind could cause a baroclinic set-down of the interface in the western end of the Mediterranean and hence lower the interface in the strait and reduce u_1. We examine each of these two possibilities in a little more detail.

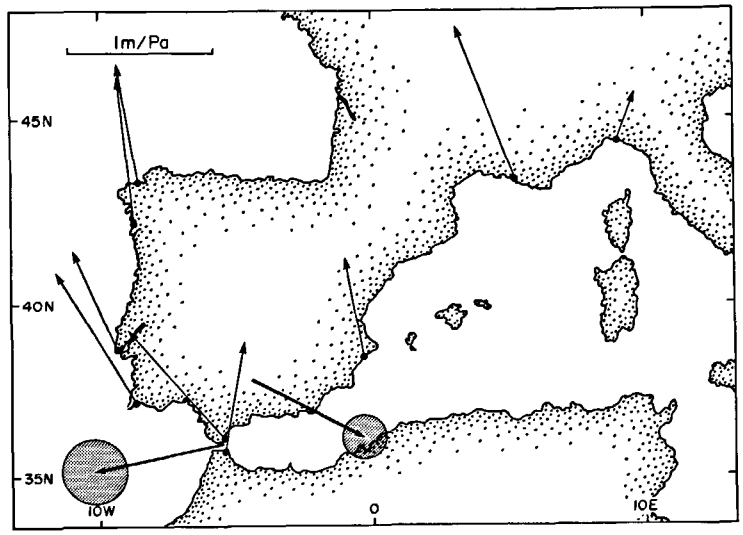

Figure 12
Wind gains for various locations (thin arrows) and for Gibraltar-Ceuta and Alicante-Lagos, using the wind halfway between each station pair (thick arrows). The shaded region denotes the standard error for the latter gains.

Table 3
Amplitude and phase of various annual cycles. Zero phase corresponds to 1 January.

	Amplitude	*Phase Lag*
$\Delta_x\eta'$ (*cm*)		
Total pressure	2.5	260°
Wind contribution	1.1	20°
Residual	3.2	242°
$\Delta_y\eta'$ (*cm*)		
Total pressure	1.1	275°
Wind contribution	1.4	204°
Residual	1.5	339°
h_1' (*m*)		
Total pressure	8.6	263°
Wind contribution	9.4	192°
Residual	10.5	321°

Upwelling in the Gulf of Cadiz

The amount of interfacial upwelling that would be produced at the head of a two-layer rectangular gulf, by a wind stress τ along the gulf axis, is independent of the earth's rotation and given by (Csanady, Scott, 1974),

$$\Delta h_1 = \tau L/(g(\rho_2 - \rho_1)h_1) \qquad (10)$$

where L is the length of the gulf.

For the Gulf of Cadiz we might take L = 100 km, ρ_2-ρ_1 = 1 kg m^{-3}, h_1 = 100 m and τ = 0.02 Pa (corresponding to the amplitude of the annual cycle in the E-W wind stress). Hence Δh_1 = 2 m, which would not change u_ρ by more than 1% and is negligible (Fig. 9). Thus, although short-term upwelling events are occasionally observed in the Gulf of Cadiz (Fiúza, 1983) and along the Moroccan Atlantic coast (Gascard, Richez, 1985), we do not expect significant seasonal

changes in the density of the inflowing Atlantic water due to upwelling. This is borne out by our analysis of hydrographic data. These data show a seasonal variation in the Gulf of Cadiz which is identical to that further away and attributable to the seasonal variation in the surface buoyancy flux (Pattullo *et al.*, 1955).

Changes in interfacial depth in the Mediterranean

We have also used (10) to estimate changes in interface height at the western end of the Mediterranean. If we now assume L = 500 km (half the length of the western basin), $h_1 = 100$ m, $\rho_2 - \rho_1 = 2$ kgm^{-3}, we obtain $\Delta h_1 = 5$ m. This in turn, via the Farmer and Armi (1986) hydraulic theory and assuming the potential control at the contraction to be flooded, would cause an increase of 1.2 cm in $\Delta_y \eta'$. As this 1.2 cm change is for 0.02 Pa, the wind gain for $\Delta_y \eta'$ is 0.6 m Pa^{-1}, not much less than the observed 0.8 m Pa^{-1}.

We note that for this very simple model the surface set-up which causes a sea level difference between Alicante and Lagos is 1.3 cm for the parameter values chosen. This can then probably explain the difference between the wind gains for $\Delta_x \eta'$ and $\Delta_y \eta'$ discussed earlier, supporting our suggestion that $\Delta_x \eta'$ is not, in fact, a good measure of just the sea level difference from one end of the strait to the other.

In the previous section we interpreted the seasonal variation of the surface inflow in terms of changing interface height in the Mediterranean and hence in the strait itself. Noting that the relationship between u_1 and h_1 is approximately linear (Fig. 7), we now attribute part of this change to the baroclinic set-up and set-down associated with seasonal changes in the eastward wind stress. The wind contribution to the change in interface height has been obtained as follows. We have calculated the seasonal variation in $\Delta_y \eta$ due to the wind (Fig. 12) and then, using the geostrophic relation (to obtain u_1) and Figure 7, we have obtained the wind induced seasonal variation in h_1 at the Sill. We have also subtracted the wind contribution from the original h_1 variation (Fig. 10) to obtain the residuals shown in Figure 13. Table 3 summarizes the amplitude and phase of the annual cycle of the interface height at Camarinal Sill from the total pressure signal, the wind contribution and the residuals. The wind contribution to the change in interface height is comparable with that estimated from the sea level data, but out of phase. The residual change in interface height, with the interface highest in April, has an annual cycle with an amplitude of about 10 m. We shall discuss in the next section the way in which this may be associated with changes in the volume of the reservoir of outflowing intermediate and deep waters.

RESIDUAL CHANGES IN INTERFACE HEIGHT

The annual cycle in upper layer thickness at Camarinal Sill (10 m, 321°) translates, via the Farmer and Armi (1986) theory, into an annual cycle in the Mediterranean reservoir (9 m, 321°). The whole seasonal variation, for which the above is the annual component, is shown in Figure 13. We note that this shows a total range of 30 m between the minimum in April (184 m) and maximum in December (214 m).

We hypothesize that the residual change in interface height is attributable to a partial draining of the reservoir of dense outflowing water during the summer when it is not being formed. The outflow consists mainly of Levantine intermediate water (Wüst, 1961) but also of Western Mediterranean deep water (Sankey, 1973; Bryden, Stommel, 1982). The interface drop is comparable to the 14 m by which the interface over the whole Mediterranean (of area 2.5×10^{12} m^2) could be lowered by the outflow of 1.7×10^6 m^3 s^{-1} (Bethoux, 1979) for a period of 8 months.

It is thought that the formation of these two types of water is localised in the center of cyclonic circulations subject to the effect of strong cold and dry winds. The formation generally occurs in late February and early March (the Levantine intermediate water and the Western Mediterranean deep water are formed in the Eastern Mediterranean basin (Morcos, 1972; Ovchinnikov, 1984) and near the South of France (Sankey, 1973; Gascard, 1978) respectively). The resulting replenishment of the reservoir of outflowing water would then spread at the speed of an internal gravity wave, leading to a raising of the interface near the strait within a few weeks.

On the basis of the above interpretation we might expect a "saw-tooth" pattern for the residual h_1, with a sudden raising of the interface in about March and a gradual lowering of the interface over the rest of the year.

The residual h_1 does conform to this pattern (Fig. 13); the raising of the interface definitely occurs more abruptly than the lowering though spread over four rather than one or two months. This smoother response may be associated with the spreading of the outflowing water, variability in the time of formation or partial replenishment of the outflowing water somewhat earlier in the year than has been previously appreciated. However, given the uncertainty in our data and the simplicity of our statistical treatment and models, it is clearly inappropriate to speculate further.

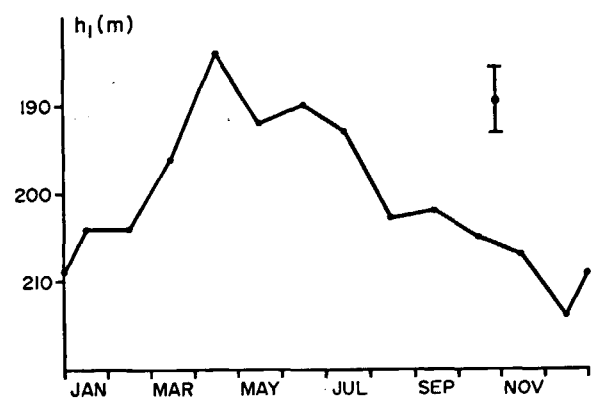

Figure 13

Seasonal variation in the interface depth over Camarinal Sill inferred, via the hydraulic control theory, from the residual sea level difference between Gibraltar and Ceuta and from density data. The standard error is mainly due to the $\Delta_y \eta$ residual.

DISCUSSION

We have investigated the seasonal variability of the surface inflow through the Strait of Gibraltar using historical data, idealised models and simple physical arguments. We are fully aware that the data base which forms the foundation of our work is of variable quantity and quality and that many of our arguments are speculative. Our final conclusions must thus be regarded as hypotheses to be checked by further field work, rather than as convincing explanations of well established facts.

Our main, directly testable, prediction is that the thickness of the inflowing layer at GC undergoes seasonal changes, with a range of about 20 m, and is associated with similar changes in the height of the interface above Camarinal Sill. Admittedly this quantitative prediction is based on the use of an idealized two-layer quasi-steady model in which the diffuse nature of the pycnocline and the existence of barotropic fluctuations (associated with tides and meteorological forcing) are neglected. Nonetheless, it is clear that suitable averaging of current meter and other data obtained over a year or more is capable of revealing changes of the depth where the mean current is zero or the density gradient a maximum.

One year or more of temperature and salinity data from the strait would also help us test our hypothesis, based on hydrographic data from near but outside the strait, that the density of the outflowing water does not change significantly and that the inflowing water shows seasonal variations that correspond to those further away in the Atlantic and are attributable to changes in buoyancy flux.

In view of the demonstration by Farmer and Armi (1986) of the significance of barotropic fluctuations in the hydraulics of two-layer exchange over a sill, it might be argued that we should not have based our discussion of the seasonal variability on the use of a steady model at all. However, for the known magnitude of the tides, and estimated magnitude of the low-frequency barotropic fluctuations due to forcing by atmospheric pressure (*see* Garrett, Majaess, 1984; Garrett, 1985), the Farmer and Armi (1986) model suggests an increase in exchange of only about 10-20% over that calculated for steady flow, so that a quasi-steady model seems quite appropriate as a first step. Moreover, even if a more elaborate model is eventually thought necessary and appropriate, it seems likely that the change in exchange, due to changes in density difference of the two layers or interface height in the Mediterranean, will be qualitatively, and perhaps even quantitatively, much as we have estimated.

Of course, if the barotropic fluctuations do affect the exchange significantly more than the 0 (10-20%) estimated above, it is possible that the seasonal variations in average surface flow, and hence exchange, implied by the sea level data are a manifestation of the seasonal variations in the storminess and hence in the variance of the barotropic fluctuations. This possibility may require further investigation, but we remark at this stage that the storminess peaks in winter and so cannot produce an annual cycle in the inflow that is in phase with either the observed or residual flow.

This leads to a discussion of our second main conclusion, that the residual changes, after the removal of the effect of wind, correspond to a partial draining of Levantine intermediate and Mediterranean deep water during the seasons when they are not being formed, and to a fairly rapid replenishment during the winter. This concept, and indeed the total inferred seasonal variation in interface height including the effect of wind, imply that the overmixing and maximal exchange assumptions of Bryden and Stommel (1984) and Farmer and Armi (1986) may only be relevant, if at all, to the later winter months, and that at other times of the year the exit control at the narrows near Gibraltar and Ceuta may be flooded and the exchange submaximal.

We have assumed in our analysis that the exchange is a maximum in March when the inferred interface height is at its highest point above the sill, but it is possible that the condition of maximal exchange is *never* achieved, and that the interface height above the sill is actually lower than implied by Figure 10, though still with approximately the same range over the course of a year. Again, this is an issue that may be settled by sufficiently long data sets from the strait itself.

It is also important to recognize that submaximal exchange would be associated with subcritical flow between Tarifa Narrows and the GC section. We have shown that existing data do indeed suggest subcritical flow, though significantly modified by friction. Further field work to check this should be a high priority.

If it does transpire that mixing in the Mediterranean is not sufficiently strong to drive the system at any time of the year to the state of maximal exchange, then a prediction of the exchange, and of the salinity and density of the outflowing water, will require a more profound understanding of the mechanisms and rates of formation of the Levantine intermediate water and of any other sources of outflowing Mediterranean water.

Clearly extensive further work in the Eastern Mediterranean including the Strait of Sicily will be required for this, but it is also possible that a study of the interannual variability in historical data sets such as those we have analyzed will be useful.

Acknowledgements

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