Low-Frequency Fluctuations in the Strait of Gibraltar from MEDALPEX Sea Level Data

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(Manuscript received 21 October 1988, in final form 1 May 1989)

ABSTRACT

Thirteen months of hourly sea level data from four stations in the Strait of Gibraltar, along with appropriate atmospheric pressure data, are used to investigate tidal and low-frequency fluctuations in the strait. Apparent recording errors in the data weaken a study of the tides on their own, but low-passed and complex demodulated (at M_2) time series show that nonastronomical variations in the alongstrait M_2 sea level slope are correlated with the varying phase lag from Tarifa to Gibraltar. More importantly, regression of the low-passed sea level difference across the strait on the varying tidal amplitude suggests that the mean surface inflow is only about 10% different from what it would be in the absence of tides.

Correlation and cross-spectral analyses also show significant connections between the three time series representing the slowly varying surface inflow, subsurface pressure gradient along the strait, and atmospheric pressure gradient along the strait. Correlation between the first and third suggests a role for direct wind forcing in the strait, but the most important result of the paper is that the ratio of fluctuations in the first two series is consistent with the basic state having been one of maximal rather than submaximal exchange during the observation period.

1. Introduction

The Strait of Gibraltar has been the site of numerous oceanographic investigations since the two-layer "inverse-estuarine" nature of the flow was first proposed in the seventeenth century (see Deacon 1971). The most recent major investigation has been the international Gibraltar Experiment (1985–86) in which a wide variety of oceanographic measurements were made. Thorough analysis and interpretation of the data obtained during this experiment, and comparison with earlier work, such as that discussed by Lacombe and Richez (1982, 1984), will undoubtedly lead to greater understanding of the oceanography of the strait on a variety of time scales. It should clarify ways in which continued long-term monitoring of some variables in the strait will lead to worthwhile integral measures of changes occurring in the Mediterranean Sea on seasonal and interannual time scales.

A critical scientific issue is whether the exchange through the strait is maximal in the sense that the exchange is as great as it can be, subject to hydraulic constraints in the strait. Bryden and Stommel (1984) argue that the exchange is driven to this limit by vigorous mixing in the Mediterranean which tends to reduce the salinity difference between Atlantic and Mediterranean waters and hence, to comply with the conservation of water and salt in the Mediterranean, increase the exchange. According to the two-layer hydraulic model of Farmer and Armi (1986), a signature of this maximal exchange limit should be hydraulically supercritical flow at the eastern end of the strait, with a shallow, fast-moving, upper layer. On the other hand, submaximal exchange would have subcritical conditions in the eastern end of the strait, with a thicker and slower inflow.

Bormans et al. (1986) examined seasonal cycles in the monthly mean sea level at Ceuta and Gibraltar (Fig. 1), assuming cross-strait geostrophy in order to deduce a seasonal cycle in surface inflow. They argued that this could only be accounted for if the flow were submaximal, with seasonal variations in interface depth that they estimated using the Farmer and Armi (1986) model. Using multiple regression techniques, Bormans et al. (1986) attributed part of these variations to meteorological influence and found that the residual showed a remarkable sawtooth pattern which they attributed to wintertime replenishment of the dense outflowing water followed by a slow draining for the rest of the year.

It would seem simple enough to check this issue of maximal or submaximal exchange by measuring directly whether the inflow at the eastern end of the strait is respectively supercritical or subcritical. Earlier data on this is rather short in duration and does not provide clear support for either interpretation as it tends to lie between the two possible hydraulic states of the inviscid two-layer theory. Bormans and Garrett (1989a,b) dis-

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cuss the matter further with consideration of rotation, realistic cross sections, friction and barotropic fluctuations. These effects tend to bring the two possible states closer together, but ambiguities remain, partly because the finite thicknesses of the shear layer and density interface at the eastern end of the strait, and the time dependence associated with the tides, weaken the application of steady state two-layer hydraulic theory.

Data obtained in 1985–86 during the Gibraltar Experiment shed further light on this issue, with suggestions that the inflow changed from generally subcritical in the fall of 1985 to supercritical by the spring of 1986 (Wesson and Gregg 1988; J. Candela, personal communication). However, even if detailed analysis confirms this, we cannot easily tell whether 1985–86 was typical. It is hoped that our increasing knowledge of the oceanography of the strait, and recognition of appropriate scientific questions, will lead to further measurements that will resolve this issue completely. In the meantime, and also to provide guidance for future studies, it is highly desirable to extract as much information as possible from any relevant historical datasets.

Our purpose in this paper is to pursue the examination of sea level data. The study by Bormans et al. (1986) was limited by the small overlap of the sea level data used and was restricted to monthly means. However, as discussed earlier, analysis of the seasonal cycle led to the view that the exchange is generally submaximal, though not greatly so. Moreover, the significant influence of wind stress on the monthly mean flow fluctuations suggests the possibility that strong wind stress events on a shorter time scale might raise the interface, between Mediterranean and Atlantic water in the Western Mediterranean inside the strait, sufficiently to require supercritical surface inflow. We hypothesized that this might manifest itself as occasional changes, in the sea level difference across the strait, corresponding to flips between subcritical and supercritical flow. In a statistical sense, we felt there was a possibility that the sea level difference, after removal of tidal changes, might be bimodal in nature, corresponding to supercritical and subcritical states. We thus acquired the 13 months of hourly sea level data collected during MEDALPEX and described by Rickards (1985, 1987), and have performed various analyses of the data from Algeciras, Ceuta, Gibraltar and Tarifa (Fig. 1).

Each dataset appears to be marred by defects, some of which are not immediately apparent, that could seriously distort the results of subsequent analyses. In view of this and the implied need for better quality control in future monitoring programs that may be proposed, we will report our preliminary analyses of the sea level data in some detail in section 2.

Of course the dynamics of the flow are related to gradients in the subsurface pressure $g\rho\zeta + p_a$ (where ζ is the sea surface elevation, ρ the water density and p_a the atmospheric pressure), or adjusted sea level $\zeta^{(a)} = \zeta + p_a/\rho g$. One might expect that, on the small scale of the strait, the variations of atmospheric pressure would be negligible, but Dorman and Arnold (1986) have drawn attention to the surprisingly large atmospheric pressure gradients that can occur along the strait. We therefore acquired available atmospheric pressure data in order to make the necessary adjustment to the sea level. This is discussed in subsection 2e.

In section 3 we discuss the results of a tidal harmonic analysis of the sea level data, particularly in the semidiurnal band, though the data errors reduce the value of some analyses. The bulk of this paper is contained in section 4 where we examine low-frequency changes in the semidiurnal tides and in sea level itself and search for correlations that might have some dynamical significance. Although a clear-cut bimodal structure in the sea level difference across the strait is not apparent, we do find significant coherence between low-frequency changes in the subsurface pressure differences. The im-

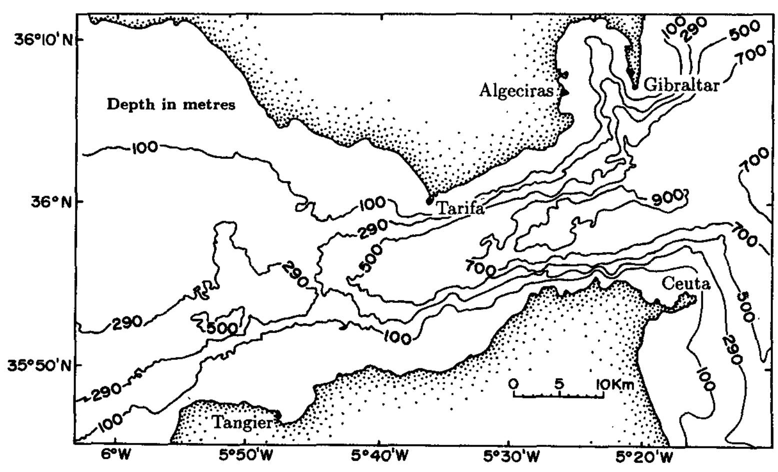

FIG. 1. The Strait of Gibraltar showing the location of places mentioned in the text.

plications of this, and of other results, are discussed in section 4 and in section 5.

2. Data, analysis techniques and preliminary results

Rickards (1985, 1987) has reported on the MEDALPEX sea level data obtained, for September 1981 through September 1982, from 29 sites in the Western Mediterranean and Adriatic. We have chosen to work only with the data from Algeciras, Ceuta, Gibraltar and Tarifa (Fig. 1). Atmospheric pressure data were obtained for Ceuta, Gibraltar, Tangier and Tarifa (see section 2e).

a. Proposed analyses

Our goal is to use differences of sea level, or adjusted seal level, to study the dynamics of the flow through the strait. As a first step we shall undertake tidal harmonic analyses of the sea level data using the package of Foreman (1977) which allows for unresolved constituents by assuming that their ratio to nearby resolved constituents is the same as in the astronomical forcing. Our analysis will be of the whole dataset (and should then give results in agreement with those of Rickards (1985)), or of shorter subsets. The latter analyses give some idea of variations in the tides, or of problems with the data. We will also study the modulation of the semidiurnal tide using complex demodulation at the M_2 frequency.

In complex demodulation the basic time series is multiplied by 2 $\cos\omega t$ and $-2\sin\omega t$, with ω the M_2 frequency, prior to smoothing with a low-pass filter. The resulting two-component vector has an amplitude and phase which can be interpreted as the slowly varying amplitude and phase of the semidiurnal tide.

The filter we use for both low-passed sea level and the tidal demodulation is a Cartwright filter of 395 points (so that we lose 147 hours at the beginning and end of each record), with a frequency response as shown in Fig. 2. It passes close to 100% of signals with periods greater than 39 h and virtually nothing with periods shorter than 28 h. When used with the tidal demodulation it passes signals within a band that is symmetric about M_2 with the same shape as for the low-pass filter at low frequencies. Note that both the low-passed sea level and demodulated semidiurnal tide are unaffected by the diurnal tides, and that the S_2 and N_2 tides are completely passed by the complex demodulation. A pure S_2 tide would thus emerge from the complex demodulation with unattenuated amplitude and phase $\delta\omega t$, where $\delta\omega$ is the frequency difference $\omega_{S_2} - \omega_{M_2} = 0.018$ rad h^{-1}.

After filtering and complex demodulation we then have, for a given input, three slowly varying outputs: the low-passed component, semidiurnal amplitude a, and semidiurnal phase lag ϕ. We then decimate these outputs to one sample per 12 hours, with 766 points

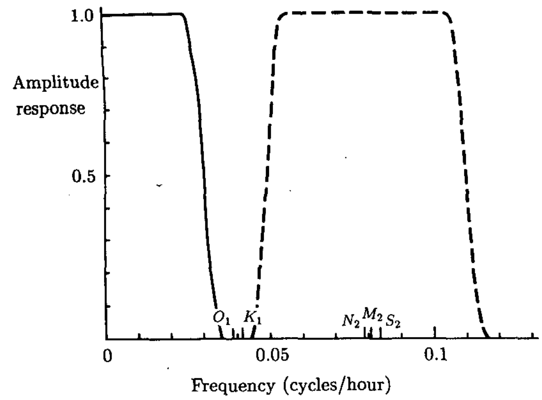

FIG. 2. Amplitude response of the low-pass filter used (solid line) and of the tidal demodulation (dashed line). The frequencies of the principal diurnal and semidiurnal constituents are shown.

(383 days) per series, before subjecting them to various statistical analyses. Before describing these it is necessary to review the data from each of the four sea level stations.

b. Gibraltar and Algeciras

The sea level record from Gibraltar contained 6 gaps of up to 14 h each. We filled these with data from nearby Algeciras (Fig. 1) since, as shown by the first part of the record for December 1981 (Fig. 3), the difference between sea level at Algeciras and Gibraltar was typically very small compared with the sea level variations at either station.

The record for the end of December 1981 and January 1982, however, shows much larger tidal differences between the two stations. The low-passed sea levels, $\bar{\zeta}$, and demodulated tidal amplitude a, coincide for most of December 1981, but differ after that in a manner that suggests an offset of one of the records by about a day with respect to the other (Fig. 3). This suggests that the data for one of the stations have been misreported. Identification of the bad dataset can be pursued by looking for unexpected changes in the tidal phase at one of the stations. We thus generated, for each station, a time series of the predicted sea level based on harmonic analysis of six months of data (November 1981 to April 1982) and compared the complex demodulated semidiurnal amplitude and phase of this with those of the raw data. The phase difference at Algeciras jumps in late December while that at Gibraltar does not, suggesting that the faulty dataset is that from Algeciras. In fact the increase in phase shift is 23°, corresponding to the 50 minutes that M_2 is delayed from one day to the next. Taken with the apparent one day offset in mean sea level and tidal amplitude shown in Fig. 3, we conclude that the Algeciras data have, for a brief period, been advanced by one day. Similar periods of suspicious Algeciras data occur later in February 1982.

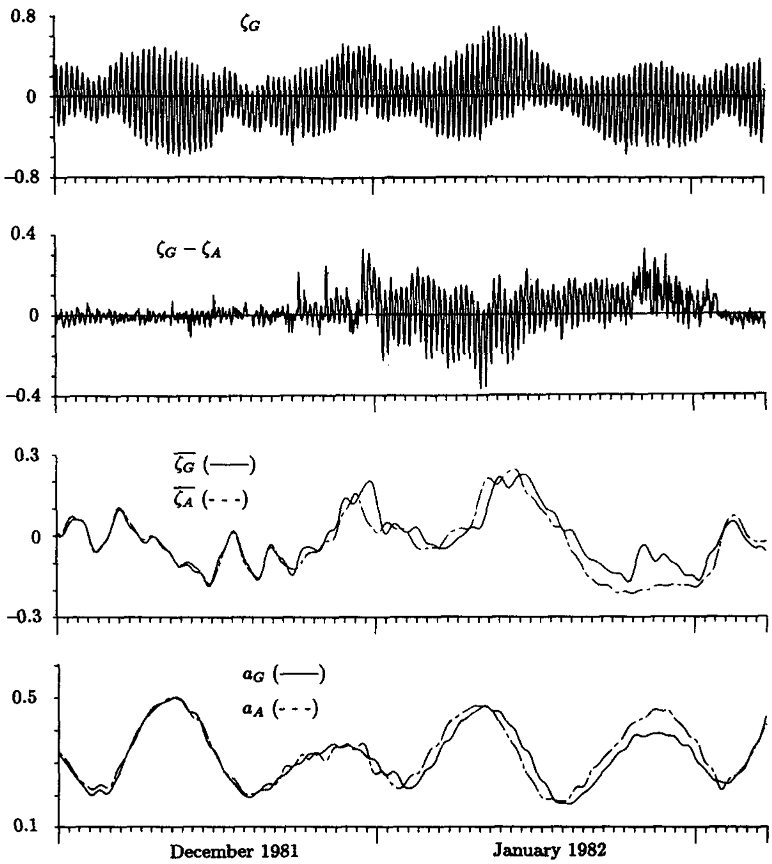

FIG. 3. Gibraltar sea level, Gibraltar sea level minus Algeciras sea level and the low-passed sea level $\bar{\zeta}$ and demodulated tidal amplitude a for the two stations, all for December 1981 and January 1982 and given in meters. (Subscripts G and A denote Gibraltar and Algeciras, respectively.)

In summary, the sea level data from Algeciras are either faulty or in agreement with Gibraltar. Fortunately the small gaps in Gibraltar are spanned by periods of agreement between the two stations. Hence, after filling the Gibraltar gaps with Algeciras data we discard the latter.

c. Ceuta

In Fig. 4 we show some results from tidal harmonic analysis of successive two-month blocks of sea level data from Ceuta and Gibraltar. The results are presented as a change in time lag $\Delta t = \Delta\phi/\omega$, where ω is the constituent frequency and $\Delta\phi$ the phase lag with respect to that computed from the first two months. After about February 1982, both M_2 and M_4 at Ceuta show a downward trend in $\Delta\phi/\omega$ until about August 1982. The values of Δt are roughly the same for M_2 and M_4, in spite of the factor of two difference in frequency, suggesting that they were both affected by the same decrease in time lag $\Delta\phi/\omega$. This could have arisen if the clock at Ceuta was running slow, being about 1

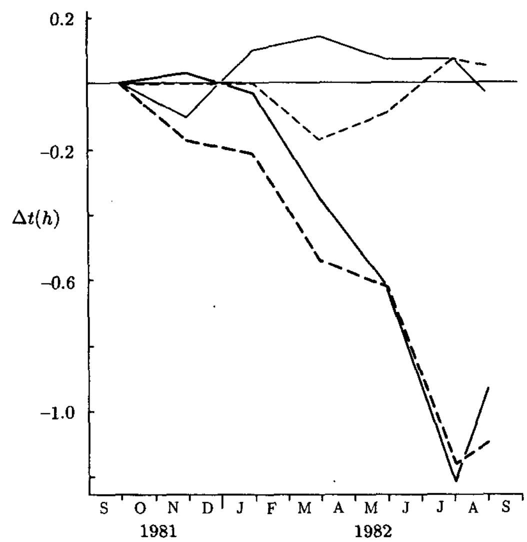

FIG. 4. Phase lag (expressed as equivalent time lag $\Delta t = \Delta\phi/\omega$) for M_2 (solid line) and M_4 (dashed line) for Ceuta (bold) and Gibraltar, based on harmonic analysis of successive two-month blocks of data.

hour behind real time by August 1982. As Gibraltar, and other stations, show no comparable changes we take this to be the correct explanation.

The Ceuta data could be corrected for this apparent clock error. However, the low-passed sea level is virtually unaffected by an error of less than an hour or so, and some aspects of the semidiurnal tides (to be discussed in section 3) are very sensitive to uncertain details of the correction and so would not be reliable even after correction. We have not, therefore, made any correction.

d. Tarifa

Although the raw data for Tarifa (Fig. 5) and the other stations show no obvious jumps in level at any point, the low-passed sea level differences between Tarifa and other stations show a jump of about 0.1 m on 28, 29 March 1982 (Fig. 5). It is clearly important to establish whether this is a real effect or another error in the data. To investigate this we examined a block of Tarifa data for 10 March to 15 April, spanning the jump, performed a tidal harmonic analysis, and calculated the residual sea level defined as the observed minus predicted sea level. Figure 6 shows this residual for 28, 29 March, with the largest jump occurring between 0600 and 0700 UTC 29 March. This jump appears as the negative outlier in the histogram of first differences of the residual sea level for 10 March to 15 April (Fig. 7). The change in recorded sea level at Tarifa thus occurs mainly in the course of one hour and is anomalous with respect to the rest of the record. It seems very unlikely that the actual sea level should change in this manner, so we ascribe the apparent shift to error.

Determining the appropriate correction for Tarifa is difficult as there is a suspicious rise in sea level during the hour before the drop. In the absence of definite information on the cause of the error we have corrected the Tarifa measurement by increasing it after March by 91 mm in order to equate the mean sea level for the last 6 months of the data to that for the first 6 months. An alternative might have been to assume no change in the difference in average sea level between Tarifa and Gibraltar from the first 6 to the last 6 months. This would have required a correction of 98 mm rather than 91 mm at Tarifa. The nature of the correction will not, in fact, significantly affect the analyses to be reported later in this paper, but we remark that an error of this magnitude seriously limits the value of data from Tarifa in any study of long term oceanographic changes that may have associated mean sea level signals of no more than a few millimeters.

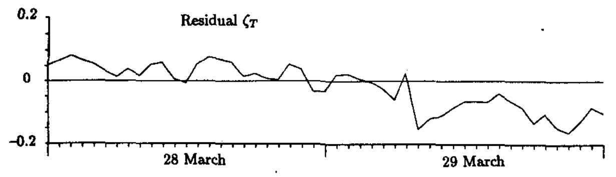

FIG. 6. Residual (observed minus predicted) Tarifa sea level in meters for 28 and 29 March 1982, based on an analysis of the data from 10 March to 15 April.

e. Atmospheric pressure data

We acquired atmospheric pressure data, for September 1981 to September 1982, for Gibraltar, Tangier and Tarifa (Fig. 1). The datasets, with a basic sampling interval of three hours, were reasonably complete for Gibraltar and Tangier, with gaps easily filled by interpolation. The data return at Tarifa was only about 50% of that at the other two stations with most of the gaps occurring at night. A multiple regression showed that, where Tarifa data did exist, 81% of the variance in the fluctuating atmospheric pressure difference between Tarifa and Gibraltar could be accounted for by equat-

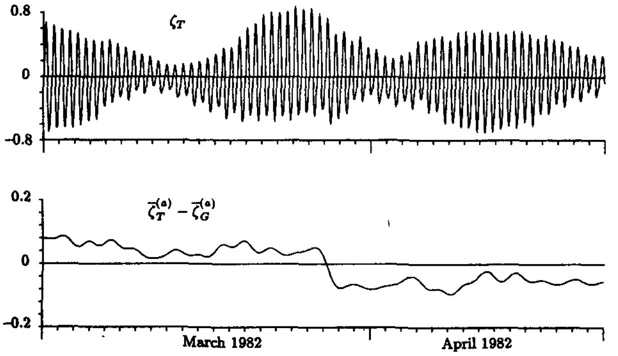

FIG. 5. Tarifa sea level and the low-passed adjusted sea level difference between Tarifa and Gibraltar for March and April 1982, in meters.

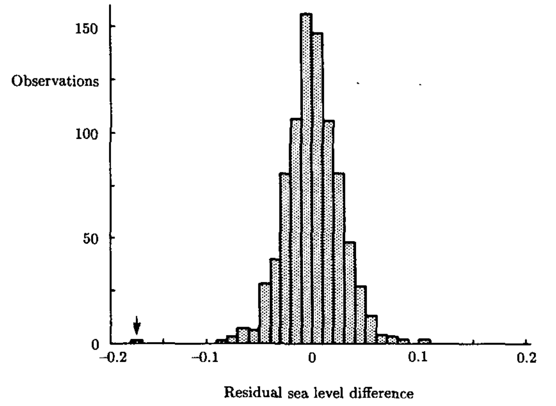

FIG. 7. Histogram of the first differences of hourly residual sea level at Tarifa for 10 March to 15 April 1982. The negative outlier is indicated with an arrow.

ing it to 63% of the pressure difference between Tangier and Gibraltar. We therefore filled the gaps in the Tarifa atmospheric pressure data using this result.

The Tarifa and Gibraltar pressure data were then further interpolated to hourly values and subjected to the same low-pass filter that had been applied to the hourly sea level data. Denoting the resulting low-passed signals by $\bar{\zeta}$ for sea level, $\overline{p_a}$ for atmospheric pressure and $\bar{\zeta}^{(a)}$ for the adjusted sea level, $\bar{\zeta} + (\rho g)^{-1}\overline{p_a}$ (with $(\rho g)^{-1} = 100$ mm/kPa), and with subscripts T and G for Tarifa and Gibraltar, we find that the root-mean-square alongstrait differences are given by

$$\mathrm{rms}\,(\bar{\zeta}_T - \bar{\zeta}_G) = 24.3,\ 24.4\ \mathrm{mm}$$

$$\mathrm{rms}\,[(\overline{p_{aT}} - \overline{p_{aG}})/(\rho g)] = 10.6,\ 11.0\ \mathrm{mm}$$

$$\mathrm{rms}\,(\bar{\zeta}_T{}^{(a)} - \bar{\zeta}_G{}^{(a)}) = 24.9,\ 20.0\ \mathrm{mm}.$$

where the first and second numbers are values for the first half and last half of the dataset. While there is only a small change in the variance of the sea level differences after adjustment, the atmospheric pressure difference has a variance equal to about 20% of the variance of the total fluctuating alongstrait subsurface pressure difference. Thus, the air pressure correction cannot be neglected.

For the cross-strait corrections we acquired atmospheric pressure data for Ceuta for January to September 1982. Its completeness was somewhat less than that of Tarifa, and it had gaps that could not usefully be filled by regression due to the lack of correlation between the difference from Ceuta to Gibraltar and the values at either Tangier or Gibraltar. However, most of the gaps at Ceuta were of 12 hours or less, so we filled gaps in $p_{aC} - p_{aG}$ by linear interpolation and found, after filtering,

$$\mathrm{rms}\,(\bar{\zeta}_C - \bar{\zeta}_G) = 37.7\ \mathrm{mm}$$

$$\mathrm{rms}\,[(\overline{p_{aC}} - \overline{p_{aG}})/(\rho g)] = 5.2\ \mathrm{mm}$$

$$\mathrm{rms}\,(\bar{\zeta}_C{}^{(a)} - \bar{\zeta}_G{}^{(a)}) = 38.0\ \mathrm{mm}$$

for the last six months of the dataset. In this case the variance of the atmospheric pressure difference is only 2% of the variance of the total pressure difference, and so is of minor importance. In view of this, and Ceuta's gappy atmospheric pressure record, we have chosen to work with the unadjusted gradient $\bar{\zeta}_C - \bar{\zeta}_G$.

3. Tides

Tidal harmonic analyses of the MEDALPEX sea level data have been presented by Rickards (1985). Her results for Gibraltar and Tarifa are hardly affected by the data gaps and spurious sea level jump respectively, but the values obtained for Algeciras and Ceuta should be viewed with caution in the light of our discussion of timing errors in the data. We do not propose to present a revised table of harmonic constants for these ports, but will discuss some results for the principal semidiurnal and diurnal constituents that have some bearing on the dynamics of the flow in the strait.

Table 1 gives values for some harmonic constants derived from the MEDALPEX dataset, using only the first four months for Ceuta (before the suspected clock problems discussed in section 2). Corresponding vectors for K_1 and M_2 are shown graphically in Fig. 8. The harmonic constants shown are for the sea level, rather than adjusted sea level, data. An analysis of the adjusted sea level for Tarifa and Gibraltar led to changes of less than 1% in the amplitude of $T - G$ and less than 2° in the phase, both of which are negligible. The lack of sufficient atmospheric pressure data for Ceuta prevented a similar study of changes in $C - G$, though there are indications of a diurnal cycle in the atmospheric pressure difference, across the strait, equivalent to a change of a few millimeters. The calculated value of K_1 for $C - G$ might thus slightly misrepresent the subsurface pressure difference at this frequency.

The diurnal tides are not well determined anyway due to their small amplitude, and $C - G$ is not well determined for the semidiurnal tides due to their near equality at Ceuta and Gibraltar and the possibility of clock errors other than those we have uncovered. Any attempt at a detailed interpretation should, perhaps, await a more reliable dataset, but for the moment we draw attention to the striking difference between $(C - G)/(T - G)$ for the diurnal and semidiurnal constituents. Is there any dynamical interpretation?

For a start, we assume cross-strait geostrophy (which should be valid at tidal frequencies provided that the cross-strait tidal currents are weak compared with the alongstrait components), and deduce the average alongstrait tidal current u' from

$$u' = g(fW)^{-1}(C - G), \tag{3.1}$$

where gravity $g = 9.81$ m s^{-2}, Coriolis frequency $f = 8.5 \times 10^{-5}$ s^{-1}, approximate strait width $W = 20$ km, and the conversion factor $g(fW)^{-1}$ is 5.8 s^{-1}. Admittedly Gibraltar is not exactly across the strait from Ceuta, so that $C - G$ may have a small contribution from the alongstrait sea level gradient. However, this contribution is likely to be much less than $T - G$ due to the widening and deepening of the strait towards the eastern end. Moreover, the harmonic constants at Algeciras (Rickards 1985) are close to those for Gibraltar, though unreliable due to the timing errors discussed in section 2.

Similarly we take $T - G$ to represent the tidal alongstrait pressure gradient

$$\rho^{-1}\partial p/\partial x = -(g/L)(T - G) \tag{3.2}$$

with $g/L \approx 6.5 \times 10^{-4}$ m s^{-1} for a distance of 16 km from Tarifa to Gibraltar.

TABLE 1. Harmonic constants for some key tidal constituents at Ceuta (C), Gibraltar (G) and Tarifa (T) with amplitude in mm and Greenwich phase lag in degrees, from tidal analysis of 13 months of data from Gibraltar and Tarifa, but only 4 months of data (September 1981 to December 1981) from Ceuta due to later timing errors there. The cross-strait ($C - G$) and along-strait ($T - G$) sea level gradients, from the differences of the complex numbers in the first three columns, and the ratio of the complex numbers $C - G$ and $T - G$, are also shown.

	C	G	T	$C - G$	$T - G$	$(C - G)/(T - G)$
O_1	19, 101°	7, 160°	6, 124°	17, 80°	4, 39°	4, 41°
K_1	38, 147°	22, 127°	26, 133°	19, 170°	5, 162°	4, 8°
N_2	66, 48°	66, 34°	89, 28°	16, 131°	24, 12°	0.7, 119°
M_2	299, 54°	316, 48°	414, 41°	36, 169°	107, 20°	0.3, 149°
S_2	113, 75°	119, 75°	158, 68°	6, −105°	42, 48°	0.1, 207°

One might hope to write the alongstrait momentum equation as

$$\partial u/\partial t + u\partial u/\partial x + \rho^{-1}\partial p/\partial x = -\lambda u. \quad (3.3)$$

We take $\partial/\partial t = -i\omega$ and write $u\partial u/\partial x$ at a tidal frequency as $u'\partial\bar{u}/\partial x + \bar{u}\partial u'/\partial x$ with $\bar{u}$ the mean, or low-frequency, flow. The momentum equation (3.3) may thus be rearranged as

$$i\omega + f(W/L)(T - G)/(C - G) = \lambda + \partial\bar{u}/\partial x + (\bar{u}/u')\partial u'/\partial x \quad (3.4)$$

if we assume that the surface tidal current at the GC section is the same as halfway between Tarifa and Gibraltar on the north shore where the pressure gradient is evaluated. The estimates of the terms on the left hand side for K_1 ($\omega = 7.3 \times 10^{-5}$ s^{-1}) and M_2 ($\omega = 1.4 \times 10^{-4}$ s^{-1}) are in s^{-1}

$$K_1\!: \quad (7.3 \times 10^{-5}, 90°) + (2.8 \times 10^{-5}, -8°) = (7.4 \times 10^{-5}, 68°) \quad (3.5)$$

$$M_2\!: \quad (1.4 \times 10^{-4}, 90°) + (3.2 \times 10^{-4}, -149°) = (2.7 \times 10^{-4}, -174°). \quad (3.6)$$

On the right-hand side of (3.4) the friction coefficient λ might be represented by $C_D\bar{u}/h$, with C_D a drag coefficient and h the water depth or upper-layer depth; most probably λ is less than 10^{-5} s^{-1} and so not significant.

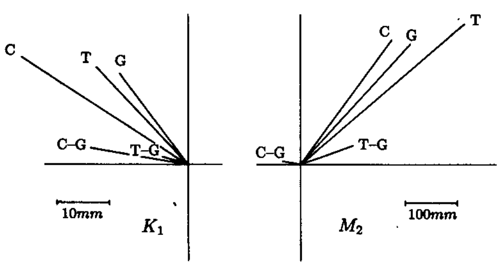

FIG. 8. Vector harmonic constants (in the form $ae^{i\phi}$, where a is amplitude and ϕ phase lag) for K_1 and M_2. Note the difference in scale for the two constituents. C, G, T refer to Ceuta, Gibraltar and Tarifa.

The term $\partial\bar{u}/\partial x$ might be negative [as required by (3.6)] if $\bar{u}$ decreases towards the east as the strait widens, but is unlikely to be more than a few $\times 10^{-5}$. The last term in (3.4) is probably similarly limited in magnitude. Thus, the right-hand side of (3.4) seems insufficient to balance the estimates of the left-hand side.

Another way of expressing this is that we expect a balance between acceleration and the surface slope at tidal frequencies, but do not find it. The lack of agreement may be associated with Ceuta, Gibraltar and Tarifa not forming an array that gives surface current and pressure gradient at the same place. In the above discussion we have assumed that the current given by $C - G$ is the same as the north shore current midway between Tarifa and Gibraltar. Alternatively, we could estimate that the alongstrait sea level difference in midstrait is $T - G + \frac{1}{2}(C - G) - \frac{1}{2}(B - T)$, where B denotes the sea level at a point on the south shore across from Tarifa. The ratio $(T - G)/(C - G)$ in (3.4) might then change by an amount of as much as $\pm\frac{1}{2}$ if, for example, the surface transport at the Tarifa section is very much less than, or twice as big as, the surface transport at the Gibraltar section. This could be a significant correction for K_1 [which has $(T - G)/(C - G) = 0.26$] but not for M_2 [which has $(T - G)/(C - G) = 3.0$]. However, in view of the poor determination of the rather small K_1 tide by the present dataset and the unreliability of $C - G$ for M_2 due to uncertainty about the timing at Ceuta, it seems inappropriate to speculate further.

It does seem, however, that $T - G$ is fairly well determined at the M_2 frequency and so, following complex demodulation at the M_2 frequency, can be used to study variations in the semidiurnal tide.

4. Tidal modulation and low frequency sea level changes

The main purpose of our study is to examine low-frequency changes in the flow through the strait. The low-passed series that we will use are $\overline{p}_{aT} - \overline{p}_{aG}$, $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$, $\bar{\zeta}_C - \bar{\zeta}_G$ (Fig. 9), with p_a the atmospheric pressure and ζ the sea level. Subscripts T, G, C indicate Tarifa, Gibraltar and Ceuta. As discussed in section 2,

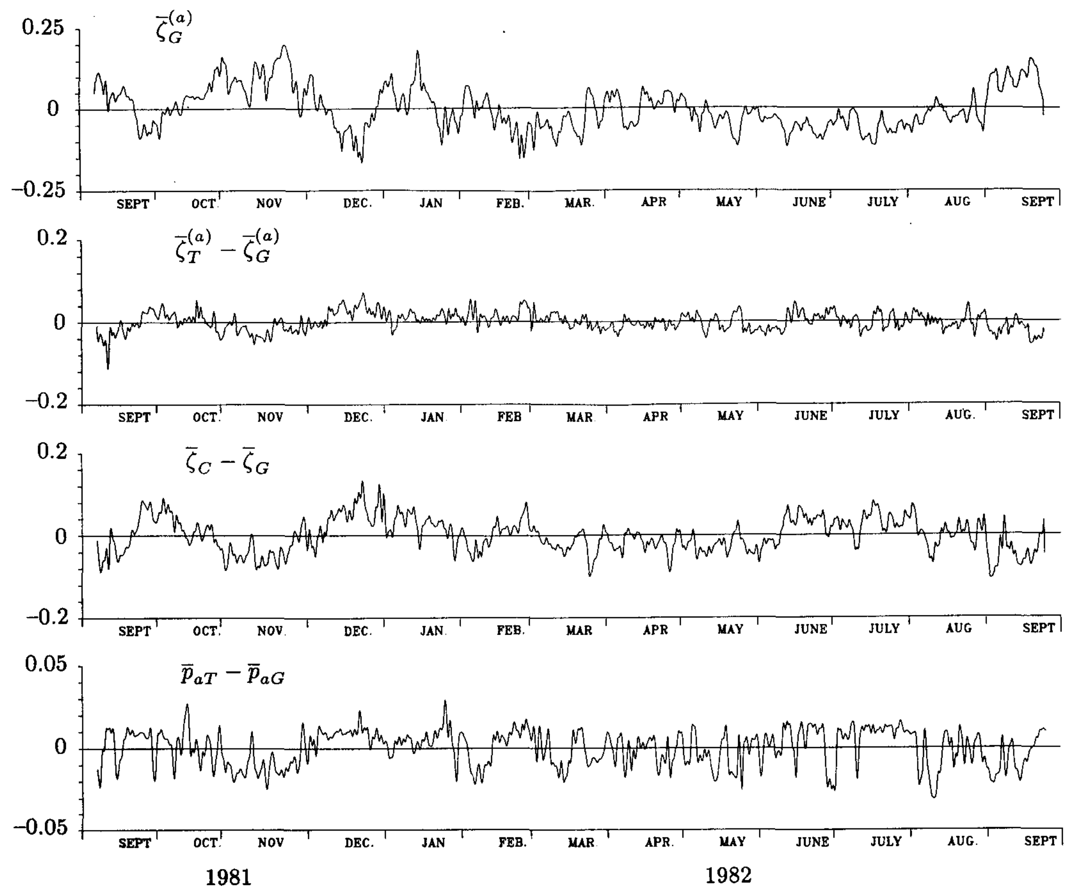

FIG. 9. Time series of low-passed datasets. All sea levels are in meters and the atmospheric pressure difference is in the equivalent unit of 10 kPa.

the alongstrait sea level gradient has been adjusted for atmospheric pressure while the cross-strait gradient has not, but the error is not significant.

Our complex demodulation provides us with time series giving the slow variation of the semidiurnal tidal amplitude a and phase ϕ at Tarifa and Gibraltar, though not usefully at Ceuta due to timing errors there. The time series that we can derive are a_T, a_G, a_{T-G}, ϕ_T, ϕ_G, ϕ_{T-G} with subscripts T and G for Tarifa and Gibraltar as before and $T - G$ for the difference from Tarifa to Gibraltar. The various quantities are connected vectorially by

$$a_{T-G}e^{i\phi_{T-G}} = a_T e^{i\phi_T} - a_G e^{i\phi_G}. \qquad (4.1)$$

There is significant modulation of these amplitudes and phases with the spring–neap and lunar apogee–perigee cycles. However, the complex S_2/M_2 and N_2/M_2 ratios are the same at Tarifa and Gibraltar, so these purely tidal cycles may be removed by normalizing with respect to one location, say Gibraltar. The four quantities a_{T-G}/a_G, $\phi_{T-G} - \phi_G$, a_T/a_G and $\phi_T - \phi_G$, connected by the formula

$$(a_{T-G}/a_G)e^{i(\phi_{T-G}-\phi_G)} = (a_T/a_G)e^{i(\phi_T-\phi_G)} - 1, \qquad (4.2)$$

then describe the way in which the semidiurnal tide varies slowly for nonastronomical reasons. We have largely restricted our attention to a_{T-G}/a_G, the normalized semidiurnal slope from Tarifa to Gibraltar, and $\phi_T - \phi_G$, minus the phase lag from Tarifa to Gibraltar, as these might provide the best indication of interaction with low-frequency fluctuations in the strait. (In fact, $\phi_{T-G} - \phi_G$ is highly correlated with $\phi_T - \phi_G$ as both are dominated, astronomical cycles apart, by variations in ϕ_G. Also, a_T/a_G is highly correlated with a_{T-G}/a_G.)

In summary, we base our analyses on the five series $(\overline{p_{aT}} - \overline{p_{aG}})/\rho g$, $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$, $\bar{\zeta}_C - \bar{\zeta}_G$, a_{T-G}/a_G and $\phi_T - \phi_G$; Table 2 shows the correlation matrix between the five variables. Due to the varying amount of autocorrelation in each time series, which affects the number of independent estimates or degrees of freedom, there is no single value for a cross correlation that is significant at the 95% confidence level. Cross-spectral analysis shows, however, that the only pairs of records that have significant coherence for periods of

TABLE 2. Correlation coefficients between the five slowly varying variables, with rms values shown on the diagonal, calculated from the whole dataset of 383 days.

	$(\bar{p}_{aT} - \bar{p}_{aG})/\rho g$	$\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$	$\bar{\zeta}_C - \bar{\zeta}_G$	$\frac{a_{T-G}}{a_G}$	$\phi_T - \phi_G$
$(\bar{p}_{aT} - \bar{p}_{aG})/\rho g$	10.9 mm	.08	.60	.15	−.21
$\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$		23.0 mm	.58	.35	−.28
$\bar{\zeta}_C - \bar{\zeta}_G$			41.7 mm	.28	−.35
$\frac{a_{T-G}}{a_G}$				.09	−.60
$\phi_T - \phi_G$					9.8°

eight days and longer are, not surprisingly, $((\overline{p_{aT}} - \overline{p_{aG}})/\rho g, \bar{\zeta}_C - \bar{\zeta}_G)$, $(\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}, \bar{\zeta}_C - \bar{\zeta}_G)$ and $(a_{T-G}/a_G, \phi_T - \phi_G)$. We start with a discussion of the connections between the different measures of tidal modulation.

a. Tidal modulation

Table 2 shows a significant correlation between $\phi_T - \phi_G$ and a_{T-G}/a_G, the time series for which are shown in Fig. 10. Cross-spectral analysis shows generally significant coherence for periods from a few days to a month. There are a few spikes in the two time series shown in Fig. 10; we do not know if they represent real physical changes or are associated with temporary timing errors at either Gibraltar or Tarifa. A scatter plot of the two time series shows that the correlation is not greatly influenced by the spikes anyway, so we have not attempted to remove them.

An explanation of the observed correlation must probably await the development of dynamical models for the barotropic and baroclinic tidal currents in the strait, though superficially it appears plausible in that a greater slope for a given tidal elevation might be associated with more influence of friction and hence more delay of the tide between Tarifa and Gibraltar. We note, however, that the phase difference reverses occasionally and, as discussed in section 3, a scale analysis of the dynamical equation for the alongstrait current suggested that friction is comparatively unimportant at the M_2 frequency. Perhaps the correlation shows a tendency for the internal tide to be larger, and travel more slowly, when the upper layer is shallow. We note that the rms fluctuation of about 10° in $\phi_T - \phi_G$ corresponds to about 20 minutes in time.

It would be interesting to examine correlations involving the modulation of the semidiurnal tidal difference between Ceuta and Gibraltar, but unfortunately the smallness of this difference, combined with apparent clock errors at Ceuta, rules this out with the present dataset.

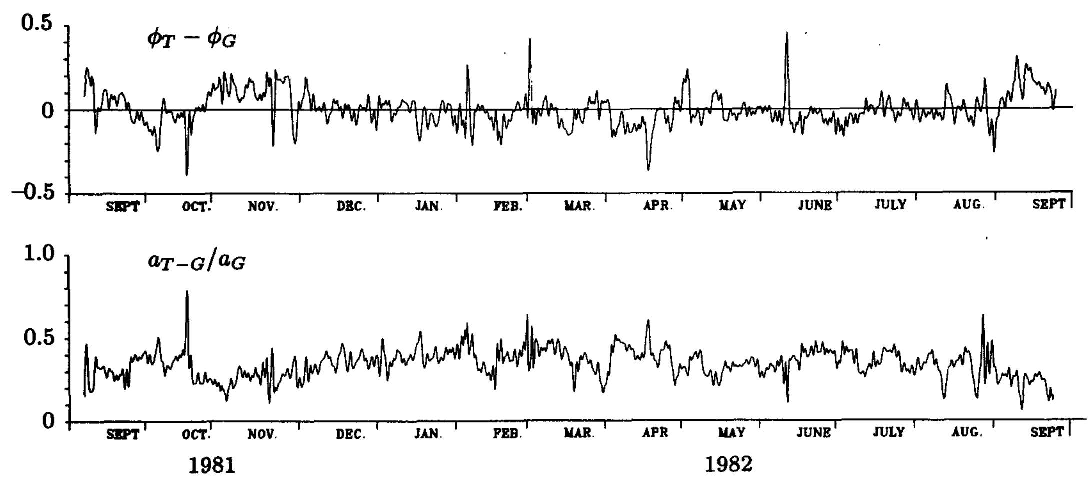

FIG. 10. Time series of the difference in M_2 phase lag (in radians) between Tarifa and Gibraltar and of the normalized tidal slope a_{T-G}/a_G.

b. Tide–mean flow interaction

Table 2 shows a slight amount of correlation between the two measures, $\bar{\zeta}_C - \bar{\zeta}_G$ and $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$, of the low-frequency flow fluctuations and our two tidal time series, though cross-spectral analysis shows no significant coherence with $\phi_T - \phi_G$ and marginally significant coherence with a_{T-G}/a_G only for periods of 16 days and longer. We conclude that, while nonastronomical modulation of the tides is presumably related to low-frequency environmental changes, we cannot associate it solely with changes in the surface current or along-strait surface slope.

Of course, a correlation between a tidal variable and a mean flow variable could also represent an effect of the tides on the mean flow, but to examine this possibility we should not remove the astronomical modulation of the tides which occurs primarily on the spring–neap (15 day) and apogee–perigee (28 day) cycles. We might, in fact, expect the mean two-layer exchange through the Strait of Gibraltar to be significantly affected by the tides as Lacombe and Richez (1982) report that at spring tides flow reversal can occur in either layer. Such an effect does not, however, appear to be very large for the surface flow as Fig. 9 shows no obvious 15 or 28 day periodicity in either $\bar{\zeta}_C - \bar{\zeta}_G$ or $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$.

To quantify this further we have regressed these two low-passed sea level series on the demodulated tidal amplitudes a_G and a_{T-G} using the entire 383 days of data. For $\bar{\zeta}_C - \bar{\zeta}_G$ the regressions are

$$\bar{\zeta}_C - \bar{\zeta}_G = 0.022 - 0.067a_G \tag{4.3}$$

$$\bar{\zeta}_C - \bar{\zeta}_G = -0.008 + 0.060a_{T-G} \tag{4.4}$$

with (4.3) suggesting that the tides decrease $\bar{\zeta}_C - \bar{\zeta}_G$ and (4.4) suggesting the opposite. Neither regression is statistically significant; the analysis will pick out not only the true tidal modulation but also a contribution from the unrelated background spectra at 15 and 28 day periods multiplied by a narrow bandwidth inversely proportional to the length of the record. Allowing for this, the regression coefficients become respectively -0.067 ± 0.042 and 0.060 ± 0.092 at the 95% confidence level. Using the M_2 values of a_G and a_{T-G} from Table 1 the regressions imply mean tidal contributions of -21 ± 14 or 7 ± 10 mm respectively, to $\bar{\zeta}_C - \bar{\zeta}_G$ or assuming cross-strait geostrophy, contributions of about -0.12 ± 0.08 m s^{-1} or $+0.04 \pm 0.06$ m s^{-1} to the average surface current at the *GC* section. The error bars do not quite allow the estimates to overlap, but, if the mean current is in fact about 1 m s^{-1} (e.g., Lacombe and Richez 1982), we conclude that the contribution from tidal rectification is no more than 10% or so. The constant terms in (4.3, 4) are, of course, meaningless in the absence of absolute leveling across the strait.

The corresponding regressions for $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$ on a_G and a_{T-G} imply tidal contributions of -3 ± 7 mm and 12 ± 5 mm, respectively, again not quite overlapping but implying an effect of no more than 10% or so if the long-term average of $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$ is about 100 mm or so, though this is not well known.

We note that, due to the energetic fluctuations at periods of a few weeks, statistical confirmation from sea level data of the small role of the tides in the mean exchange will require even longer datasets than the 13 months used here. However, our preliminary conclusion, that the tides do not significantly affect the mean surface flow through the Strait of Gibraltar, is important as it suggests that quasi-steady hydraulic models of the flow (Farmer and Armi 1986; Bormans and Garrett 1989b), which ignore the influence of the highly time-dependent tidal fluctuations, might be more valid than expected. On the other hand, Candela et al. (1989) do find significant spring–neap variability in the flows, at Camarinal sill, recorded during the Gibraltar Experiment. Further investigation and quantitative assessment is required.

c. Low-frequency fluctuations

The time series for $\bar{\zeta}_C - \bar{\zeta}_G$ and $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$ (Fig. 9) show no obvious flips between two different values. Moreover, histograms of the two datasets and the scatter plot (Fig. 11) show no evidence of bimodality. Our initial hypothesis that this might occur due to occasional changes between subcritical and supercritical inflow states thus appears to be unfounded, at least for the 13 months for which we have data.

There is, however, some correlation between the two

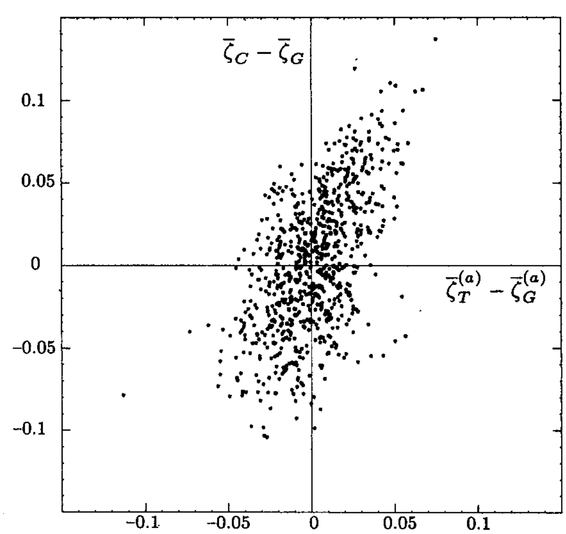

FIG. 11. Scatter plot of the low-passed sea level differences $\bar{\zeta}_C - \bar{\zeta}_G$ and $\bar{\zeta}_T^{(a)} - \bar{\zeta}_G^{(a)}$, in meters.

variables (Table 2 and Fig. 11). Before proceeding further it is convenient to abbreviate our notation to

$$U = \bar{\zeta}_C - \bar{\zeta}_G, \quad H = \bar{\zeta}_T{}^{(a)} - \bar{\zeta}_G{}^{(a)},$$
$$P = (\bar{p}_{aT} - \bar{p}_{aG})/\rho g \quad (4.5)$$

for the measures of the surface inflow (U for current), alongstrait subsurface pressure gradient (H for head) and alongstrait atmospheric pressure gradient respectively. As there is no reason to suspect relatively more noise in either U or H, an appropriate measure of the ratio of the coherent parts of the two variables is probably the slope of the major axis of the elliptical cloud of points in Fig. 11. More precisely, we rotate the axes until the new variables are uncorrelated, which is equivalent to finding the dominant empirical orthogonal function. The ratio is given by

$$R = A \pm (A^2 + 1)^{1/2} \quad (4.6)$$

where $A = (\delta^2 - 1)(2\delta\rho)^{-1}$ with δ^2 the variance ratio of U to H and ρ the correlation coefficient. With $\rho > 0$ we choose the plus sign in the formula for R to get the major axis, or dominant orthogonal function, and with $\delta = 1.81$ and $\rho = 0.58$ (see Table 2) we obtain $R = 2.6$. As discussed by Bormans and Garrett (1989b), the ratio of the fluctuating gradients across and along the strait should provide a clue to the dynamics of the flow fluctuations even in the absence of data on the depth of the interface at the eastern end of the strait.

The basic physics is that, for the widening strait from Tarifa to Gibraltar, an inviscid supercritical upper layer accelerates and hence flows downhill, giving positive H as well as U. For a flow that is controlled at Tarifa Narrows, U and H both increase with increasing volume flux, so that the fluctuations of U and H have a positive ratio.

On the other hand, a subcritical inviscid flow decelerates as the strait widens and so flows uphill, giving negative H for positive U. Fluctuations may be produced by either increasing the flow and keeping it marginally subcritical at Tarifa Narrows or maintaining the same flow rate but changing the interface height at the eastern end of the strait. In the former case, the *magnitude* of both U and H scale with the flow, so that the fluctuation ratio is negative. In the latter case the changes in interface depths are greater at the Bernoulli dimple at the narrows than at the eastern end of the strait, so again the fluctuation ratio is negative.

One might expect friction to change these results significantly, but Bormans and Garrett (1989b) point out that, from the hydraulics of a reduced-gravity model with widening of the strait and friction, the effect of widening tends to be greater than that of friction if, as is required for observations of separated flow in the strait (Farmer and Armi 1989), a control section exists near the narrows. Plausible values of an interfacial friction coefficient also imply that the role of friction is comparatively minor.

The actual predictions of fluctuations in U and H were made with a two-layer hydraulic model that allows for nonrectangular cross-sections and interfacial and bottom friction. Bormans and Garrett (1989b) also allow for the effect of the earth's rotation in estimating the alongstrait pressure gradient at the north shore rather than in midstrait. They find that $R = U/H$ should be about 1.4 for a fluctuating supercritical inflow, but large and negative for a fluctuating subcritical inflow. The value 2.6 which we have obtained is thus more compatible with supercritical flow, though larger than predicted. Moreover, the correlation between the two variables is not particularly high, suggesting the presence of noise or the influence of other inputs.

In fact, as shown by Table 2, the alongstrait atmospheric pressure difference, which may be a measure of the alongstrait wind (Dorman and Arnold 1986), is correlated with U though not significantly with H, suggesting that it is an extra variable that drives the inflow fluctuations. We shall investigate possible dynamical reasons for this later; for the moment we remove the influence of $P \equiv (\bar{p}_{aT} - \bar{p}_{aG})/\rho g$ from the other two series and reexamine them.

To do this we can use the variables $U - 2.30P$ and $H - 0.17P$ in which the coefficients are determined by separate regression of the two flow variables on P. These adjusted variables are shown in a scatter plot in Fig. 12; the correlation coefficient ρ is 0.67, the standard deviation ratio δ is 1.46 and the slope of the major axis, given by (4.6), is 1.7, closer to the theoretical value 1.4 for a fluctuating supercritical flow.

We also note that the rms values of $U - 2.30P$ and $H - 0.17P$ are 33.4 mm and 22.9 mm respectively. From the model of Bormans and Garrett (1989b), these would require an rms fluctuating barotropic flow through the strait of 0.67×10^6 or 0.62×10^6 m^3 s^{-1}, respectively. These are slightly more than the 0.4×10^6 m^3 s^{-1} or less that Garrett and Majaess (1984) and Garrett (1985) estimated might be required by continuity in response to atmospheric pressure changes in the Mediterranean, but could perhaps represent the additional response to wind-induced setup or setdown in the Mediterranean. However, 0.4×10^6 m^3 s^{-1} was actually measured by Candela et al. (1989).

Even though the residual sea level differences, with the forcing by pressure difference (or wind) removed, are still not perfectly correlated, there is no apparent signature of fluctuations on subcritical flow. According to the preferred model of Bormans and Garrett (1989b), such fluctuations would appear in a scatter plot, like that in Fig. 12, with a large negative slope and with an origin, for zero barotropic flow, offset about −0.08 m in both sea level differences from that for supercritical fluctuations. There is no sign of such fluctuations in Figs. 11 or 12.

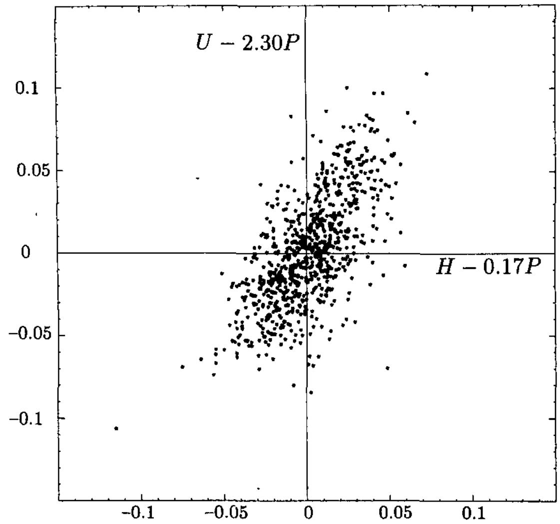

FIG. 12. Scatter plot of the low-passed residual sea level differences $(\bar{\zeta}_C - \bar{\zeta}_G) - 2.30(\bar{p}_{aT} - \bar{p}_{aG})/\rho g$ and $(\bar{\zeta}_T{}^{(a)} - \bar{\zeta}_G{}^{(a)}) - 0.17(\bar{p}_{aT} - \bar{p}_{aG})/\rho g$ after removal, by regression, of the effect of the alongstrait atmospheric pressure gradient.

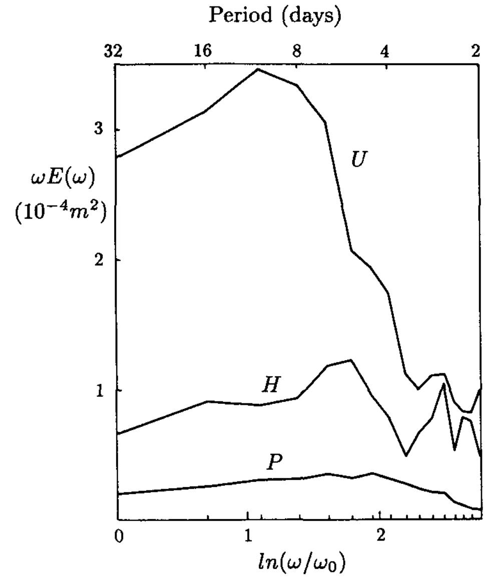

FIG. 13. Power spectra of the three series $U = \bar{\zeta}_C - \bar{\zeta}_G$, $H = \bar{\zeta}_T{}^{(a)} - \bar{\zeta}_G{}^{(a)}$ and $P = (\bar{p}_{aT} - \bar{p}_{aG})/\rho g$. The reference frequency ω_0 is $1/32$ cycles day^{-1}.

d. *The frequency-dependent response*

Our discussion of low-frequency fluctuations in subsections (4.3, 4) has depended on averages over the whole dataset. It ignores any frequency dependence of the interrelationships between U, H and P and also leaves the (admittedly unlikely) possibility that the high correlation coefficients we have calculated are merely a consequence of trends in the data and do not have any statistical significance. We have thus extended our study to the frequency domain using spectral and cross-spectral analyses.

We use the fast Fourier transform package of Carter and Ferrie (1979), in which the data are divided into n successive 50% overlapped blocks, to each of which a full cosine bell is applied to minimise leakage. This provides $18n/11$ degrees of freedom. We pad our datasets of 766 points in each series by adding a zero at each end (which is ignored anyway due to the cosine bell, and has a negligible effect on leakage). We then divide the 768 point records into 23 overlapping blocks of 64 points each, so that the longest period resolved is 32 days. The Nyquist period is 1 day, shorter than our filter cutoff period; we will restrict our attention here to periods of 2 days and longer. There are 38 degrees of freedom in our spectral estimates.

Figure 13 shows the power spectra of the three low-passed series and Fig. 14 the coherence and phase between the pairs (U, H) and (U, P); the coherence between the third pair (H, P) is generally insignificant, as expected from the correlation coefficient (Table 2), and is not shown. It is interesting that the coherence between U and P has a peak, centered on a period of

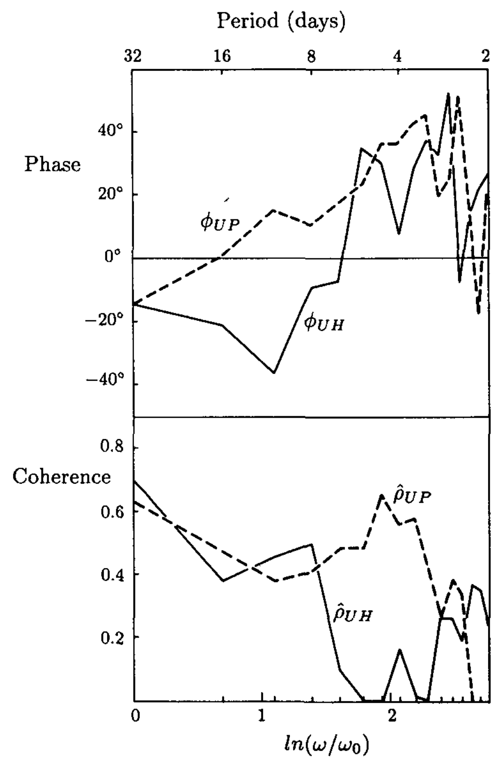

FIG. 14. Coherence $\hat{\rho}$ and phase ϕ between the pairs (U, H) and (U, P). The coherence has been corrected for bias according to the formula $\hat{\rho}^2 = \rho^2 - 2\nu^{-1}(1 - \rho^2)$ where $\nu = 38$ is the number of degrees of freedom and ρ is the computed coherence (Jenkins and Watts 1968). Corrected coherences above 0.33 are significant at the 95% level. Positive phase corresponds to U lagging the other series. Error bars on the phase lags are not shown, but are ±32° (at the 95% confidence level) for $\hat{\rho} = 0.6$.

about 5 days, where the (U, H) coherence falls off, suggesting predominantly wind forcing in that band.

As was done in the time domain, we can use regression to remove the effect of P from U and H before comparing them. Denoting the residuals by U', H', we have

$$\rho_{U'H'}e^{i\phi_{U'H'}} = [\rho_{UH}e^{i\phi_{UH}} - \rho_{UP}\rho_{HP}e^{i(\phi_{UP}-\phi_{HP})}] \times (1-\rho_{UP}^2)^{-1/2}(1-\rho_{HP}^2)^{-1/2} \quad (4.7)$$

where the coherences are as computed (i.e., uncorrected for bias). The bias-corrected coherence $\hat{\rho}_{U'H'}$ is shown in Fig. 15; we note that it is considerably greater than $\hat{\rho}_{UH}$ and is statistically significant at the 95% confidence level for periods longer than 4 days or so.

The appropriate fluctuation ratio is again given in magnitude by (4.6), with δ^2 the spectral energy ratio of U' to H' and ρ equal to $\rho_{U'H'}$. There is a tendency for this ratio to decrease with frequency (Fig. 16); it

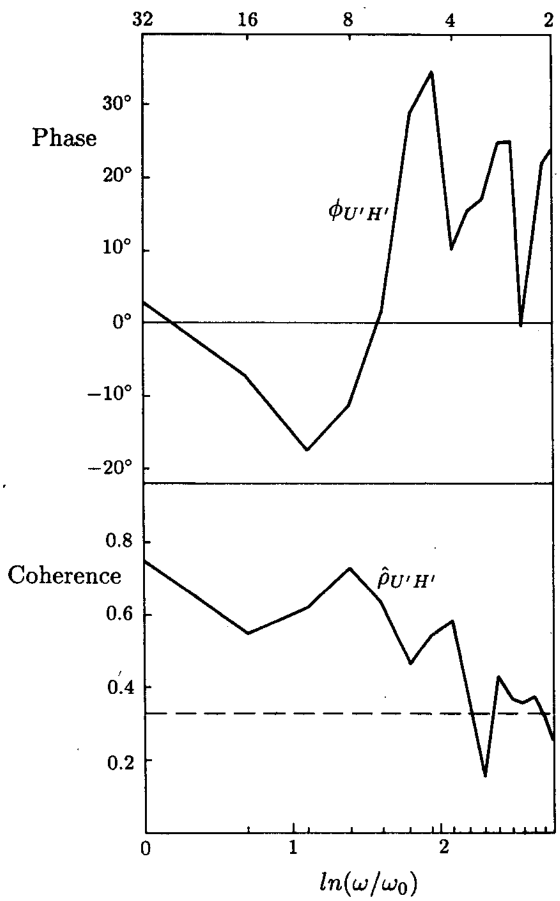

FIG. 15. Bias-corrected coherence and phase between the residuals of U and H after removal of the effect of P on both. The dashed line shows the coherence of 0.33 that is just significant at the 95% level. Positive phase corresponds to U' lagging H'.

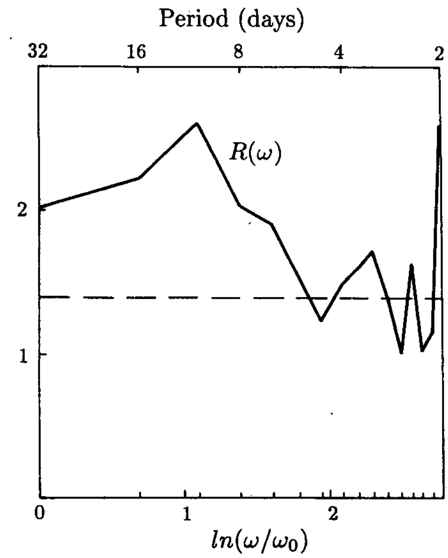

FIG. 16. Ratio of the residuals U' and H' after the removal of the effect of P. The theoretical value of 1.4 for quasi-steady supercritical fluctuations is shown by the dashed line.

is somewhat larger than the theoretical value of 1.4 for quasi-steady fluctuations.

A modification of the quasi-steady theory of Bormans and Garrett (1989b) would include time dependence in both momentum and continuity equations. Ignoring it in the latter, but including the acceleration as in (3.3), leads to a predicted fluctuation ratio

$$U'/H' = R[1 - i\omega LR(fW)^{-1}]^{-1} \quad (4.8)$$

as a function of frequency where R is the ratio at zero frequency. With width $W \approx 20$ km, distance L from Tarifa to Gibraltar about 16 km, $f = 8.5 \times 10^{-5}$ s^{-1}, and $R = 1.4$, the phase lag, which increases linearly with frequency, would have a value of 13° at a 4 day period. We have not shown, in Fig. 15, the error bars in the computed phase lag, but these are such that the phase is not significantly different from either zero or the prediction of (4.8). We also note that (4.8) would only predict a reduction by about 3% in the magnitude of U'/H' at a period of four days, so that the trend in Fig. 16 is more likely to be an artifact of the data.

e. The effect of wind

We have seen that the alongstrait atmospheric pressure gradient is significantly correlated with the surface inflow. One possible explanation might be that this pressure gradient represents the atmospheric pressure difference from Atlantic to Mediterranean which, as mentioned above, forces a net barotropic inflow, effects the two-layer exchange and modifies U. If this is the case, though, H should also be affected with a change R^{-1} times that in U. In other words, P should be just as well correlated with H as with U, which it clearly is not (Table 2).

We thus conclude that P represents an independent input as well as, possibly, driving a barotropic transport Q_0 that affects U and H. Our conceptual model is

$$U = R\alpha Q_0 + S\beta P, \quad H = \alpha Q_0 + \beta P \qquad (4.9)$$

with U and H driven by changes in Q_0 and, independently, in P. Now P could be a proxy in our dataset for the alongstrait wind. For submaximal exchange this could affect U and H by causing setup or setdown, in the Mediterranean, of the interface between Atlantic and Mediterranean waters (Bormans and Garrett 1988b). However, we have already deduced that the exchange was maximal, rather than submaximal, by examining the ratio of U to H with the effect of P removed from each by regression.

The most likely remaining scenario is that the observed effect of P on U and (negligibly) on H represents direct wind forcing of the flow in the strait. Given the observed correlation of 0.60 between U and P (Table 2), we then ask whether 36% of the variance of U can be driven by P, i.e., whether the wind in the strait can produce an rms 25 mm in the sea level difference across the strait, equivalent to an rms surface current $25 \times 5.8 \approx 140$ mm s^{-1}. A proper answer to this question requires the addition of an alongstrait wind stress to the hydraulic model of, for example, Bormans and Garrett (1989b). The output of the model would also give a theoretical value for the ratio S of U and H responses in (4.9), and make it possible to correct for the contribution, to the correlation between U and P, of the correlation between Q_0 and P.

Neglecting these considerations for the moment, we assume that the rms alongstrait wind stress is about 0.2 Pa, and estimate its influence on the alongstrait current u from the momentum equation

$$udu/dx \approx \tau/\rho h. \qquad (4.10)$$

Applying this at the eastern end of the strait with $h \approx 50$ m for supercritical flow and an alongstrait distance of 20 km, we have, taking $u \approx 1$ m s^{-1}, an increase in u of about 80 mm s^{-1}, not too much less than the 140 mm s^{-1} estimated from the data. It does thus seem possible that wind fluctuations within the strait itself can contribute to fluctuations in the surface inflow.

In the frequency domain we note that, while not statistically significant at any single frequency, the phase lag of U behind P shows a tendency to increase with frequency, with a value of about 40° at a period of 4 days. While we repeat that the coherence and phase between U and P may partially arise from the correlation of Q_0 and P in our simple model, adding an acceleration term $\partial u/\partial t$ to (4.10) would give a phase lag of U behind τ that increases with frequency and is about 16° at four days, not completely out of line with the observations.

5. Discussion

The substantial length, near-completeness and simultaneity of the MEDALPEX sea level datasets from the Strait of Gibraltar have made them invaluable in a study of the physical oceanography of the strait. However, various features of the data that might have been misinterpreted in terms of physics appear instead to have been measurement errors. While gaps at Gibraltar, misreporting at Algeciras and a datum shift at Tarifa have been corrected for in this study and have not affected our analyses, an apparent clock error at Ceuta has seriously weakened our investigation of the tidal regime as the M_2 sea level difference between Gibraltar and Ceuta is highly sensitive to phase errors at either station.

Nonetheless, we have found an interesting correlation between the varying M_2 slope from Tarifa to Gibraltar and the phase lag between these two stations. The result awaits explanation in terms of a proper model of the barotropic and baroclinic tides in the strait. We emphasize that the tides have significant nonastronomical low-frequency variability and a plausible model of them must account not only for the average tide, but also for relationships, like the one we have found, between features of the variability. In view of this variability, we also stress the value of basing tidal studies on simultaneous sea level records from different stations rather than on harmonic constants from records obtained at different times.

The dominant variability in the semidiurnal, or diurnal, tides as a whole still comes from the beating of different constituents, rather than from nonastronomical modulation of individual constituents. In particular, S_2 and N_2 cause significant 15 and 28 day modulations of the semidiurnal tide dominated by M_2 but, somewhat surprisingly in view of the strength of the tidal currents, this does not lead to noticeable 15 and 28 day periodicity in the low-passed sea level differences which we use as measures of the sub-tidal surface flow. Our statistical analyses have suggested that the presence of a semidiurnal tide in the strait changes the mean surface flow by no more than about ±10% from what it would be with no tides. Refining this estimate with sea level data will require substantially more than the 13 months of data used here. It is possible that shorter records of other types of data (e.g., from current meters) could establish the magnitude of the effect (e.g., Candela et al. 1989), but a fundamental limitation arises from the high energy level of broad band fluctuations due to other causes.

When it comes to low-frequency variations per se, it is clear that, in the alongstrait direction at least, the atmospheric pressure gradient is an important part of the total pressure gradient, so that adjusted, rather than raw, sea level data is required. We have then found that the ratio of adjusted sea level fluctuations across and along the strait shows strong evidence that the two-layer exchange was maximal for the observation period.

The fluctuations are thought to be largely driven by the fluctuating barotropic transport through the strait, though the data also suggest a role for the local wind. Separating out the effects of the two forcing mechanisms requires extension of the modeling efforts of Bormans and Garrett (1989b) to include local wind forcing.

The theoretical predictions so far have been based on steady state models, whereas the fluctuations are, of course, time-dependent. The corrections are small, but theoretical estimates of them imply phase lags between different datasets that are not inconsistent with the results of our statistical analyses.

The fact that we have been able to draw significant conclusions from sea level and atmospheric pressure data alone does not reduce the need for independent measurements of currents, the interface height and other flow properties. It does, however, suggest that sea level measurements can be a cheap and effective component of future studies and monitoring. In particular, it would be exciting if it could be confirmed that the ratio of adjusted sea level differences across and along the strait, after removal of the effects of wind, really can discriminate between maximal and submaximal exchange.

This question, of maximal or submaximal exchange, remains in our opinion the most important issue in the physical oceanography of the Strait of Gibraltar. We have shown that the fluctuations in adjusted sea level differences at the time of the MEDALPEX project imply that the exchange was maximal then, but there is also evidence (e.g., Bormans et al. 1986; Wesson and Gregg 1988) for submaximal exchange at other times. This suggests that changes between the two states occur, with the change from maximal to submaximal exchange presumably associated with flooding of the hydraulic control near Tarifa, by a lowering of the interface between Atlantic and Mediterranean waters in the Western Mediterranean (Farmer and Armi 1986; Bormans and Garrett 1989b). Further work is required to establish the reasons for this and the time scales on which it occurs.

Acknowledgments. We thank Leslie Rickards for supplying the MEDALPEX sea level data, and the Spanish and British organizations responsible for the original measurements. We also thank the U.S. National Climatic Center and Abdelaziz Belhouji for providing us with atmospheric pressure data. We are grateful to Myriam Bormans and Julio Candela for discussion of the issues addressed in this paper, and to other colleagues in the Gibraltar Experiment for general discussion. Jackie Hurst provided excellent assistance in the preparation of the paper. Our work is supported by the Natural Sciences and Engineering Research Council and by the Office of Naval Research through Contract N00014-85-G-0098.

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