Nonisostatic Response of Sea Level to Atmospheric Pressure in the Eastern Mediterranean

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(Manuscript received 30 September 1983, in final form 11 January 1984)

ABSTRACT

We analyze 5 months of sea-level data from Katakolon, Greece, in terms of local atmospheric pressure and the two components of geostrophic wind. The response to pressure is isostatic at low and high frequencies, but significantly nonisostatic for intermediate frequencies centered on about 0.01 cycles per hour. The response is consistent with a simple theory in which the fluctuating barotropic flow through the Straits of Gibraltar and Sicily is geostrophically controlled at low frequency. The local geostrophic wind contributes very little to the sea level variance; the response coefficients, while not well determined, are qualitatively as expected and quantitatively correspond to a very narrow near-shore region of shallow water.

1. Introduction

The response of the Mediterranean to atmospheric pressure is limited by the Strait of Gibraltar. At sufficiently low frequency there is time for water to escape through the Strait and permit an isostatic response to atmospheric pressure in the Mediterranean (assuming that this occurs in the Atlantic), whereas at some sufficiently high frequency the constriction of the Strait prevents an isostatic response.

It is of some scientific interest to establish, both observationally and theoretically in terms of models for flows through straits, just what is the sea level response to changing atmospheric pressure. The results for sea level itself are probably not of any great practical consequence, but they are closely related to the fluctuating flow through the Strait of Gibraltar. Lacombe (1961) pointed out that an isostatic response of the Mediterranean would require, by continuity, a fluctuating component of the flow through the Strait of Gibraltar of the order of 0.5 m s^{-1}, comparable with that observed and comparable also with the mean and tidal flows in the Strait. The surface inflow, or bottom outflow, in the Strait, averaged over a few tidal periods, can thus vary very substantially and cause low-frequency variability in oceanographic conditions on either side of the Strait. Cheney and Doblar (1982), e.g., suggested that the extent of the anticyclonic gyre in the Alboran Sea (Fig. 1) depends on the varying inflow at the Strait of Gibraltar, and hence on the large-scale meteorological conditions affecting that variability.

Garrett (1983) discussed the response of the Mediterranean to varying atmospheric pressure, drawing attention to the importance of the Strait of Sicily as well as the Strait of Gibraltar. He pointed out that the flow through the Straits, in response to sea level differences between the ends, is controlled at low frequencies by the requirement for a geostrophically balanced slope across the Straits (see Garrett and Toulany, 1982; Toulany and Garrett, 1984). The critical dimensionless parameters are $\epsilon_i = \omega f A_i (gH_i)^{-1}$ for $i = 1, 2$, where ω is the frequency of the fluctuating flow, f the Coriolis frequency, A_1, A_2 are the areas (approximately 8.4×10^{11} m^2 and 1.7×10^{12} m^2, respectively) of the western and eastern basins and H_1, H_2 the mean depths, at the sills, of the two straits.

For very small values of ϵ_i (low frequency) the flow through the Straits is sufficient to permit an isostatic response of the Mediterranean basins to atmospheric pressure (or to sea level change in the Atlantic due to other causes). If the parameters ϵ_i are not very small, there is insufficient time for the flow through the Straits to equalize the sea levels between the Atlantic and Mediterranean. Thus, even for an isostatic response in the Atlantic outside the Strait of Gibraltar, one would expect a nonisostatic response of the Mediterranean. Garrett (1983) evaluates the departure from isostasy in terms of a simple model that treats the pressure variations as traveling disturbances, and points out that the effect should be particularly evident in the eastern Mediterranean.

In fact, as one proceeds to higher frequencies the Straits prevent equalization between sea levels in the different areas, but the spatial scales of the atmospheric pressure become comparable with, or less than, the scales of the basins, and local isostasy is still possible because of purely internal adjustments.

For plausible values of the parameters (with H_1, H_2 being the only ones that are rather uncertain) the theory of Garrett (1983), and its extension in Section 4 of this paper, suggest that a nonisostatic response of the Eastern Mediterranean should be evident for the pe-

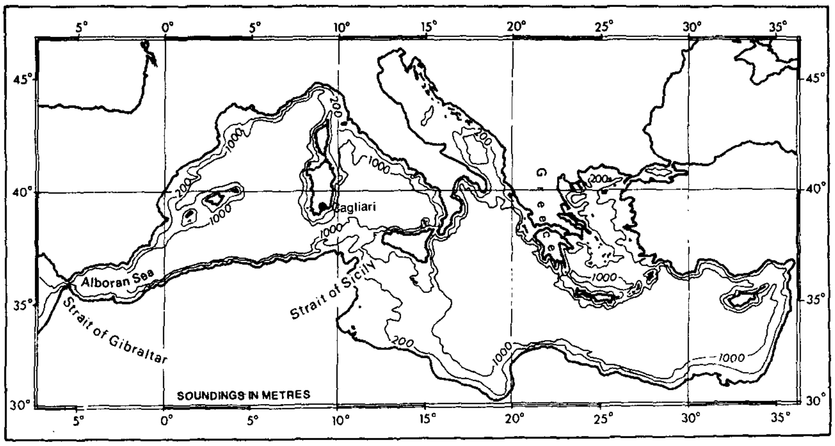

FIG. 1. The Mediterranean Sea, showing the 200 and 1000 m isobaths.

riods of several days which contain much of the meteorological variability. In view of this, sea level and meteorological data from a suitable location in the eastern Mediterranean were sought. The data obtained are described in Section 2. In Section 3 the statistical analyses employed are discussed and the results, showing the connection between sea level and meteorological forcing, are presented. The relevant theory is discussed in Section 4 and the comparison of observations and theory is made in Section 5. The paper concludes in Section 6 with a discussion of the results and their implications.

2. Data

The basic datasets used in this study are:

1) Hourly sea level heights for Katakolon, Greece (Fig. 2) for the period of 150 days from 0100 LST 1 January 1982 to 2400 LST 30 May 1982.

2) Hourly atmospheric pressure from Kerkira, Andravidha and Samos (Fig. 2) for the same period. (In fact, in view of our predominant interest in low frequencies and the slowly changing nature of atmospheric pressure, we used values of atmospheric pressure every 12 h, with a linear interpolation to hourly values).

The atmospheric pressure for Katakolon was assumed equal to that for Andravidha. All three atmospheric pressure datasets were used to define, at any instant, the coefficients in the formula

$$P_a = Nx - Ey + P, \tag{2.1}$$

where P is the pressure at Andravidha (taken as the origin), x is positive towards the east and y towards the north. The coefficients N and E are proportional to the northward and eastward geostrophic wind, and hence may be used in the multiple regression of Ka-

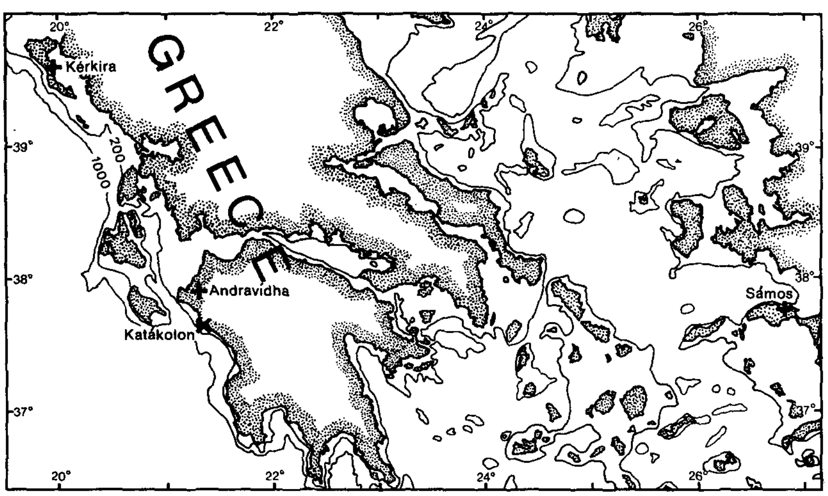

FIG. 2. Location of the sea level station (×) and atmospheric pressure stations (+).

takolon sea level on atmospheric pressure and wind (see Section 3).

Figure 3 shows the Andravidha atmospheric pressure data and the Katakolon sea level data after treatment with the $A_{24}{}^2A_{25}/(24^2 \times 25)$ tide-killing filter (Fig. 4). The data suggest a sea level response to atmospheric pressure that is close to being isostatic.

3. Data analysis

a. Techniques

The data shown in Fig. 3 have been filtered to remove the tides from the sea level data. However, the rms tidal amplitude at Katakolon is only about 50 mm, so that leakage from the tidal bands into lower frequencies is not a problem in spectral analysis of the raw data. We thus work with unfiltered time series of each variable, with hourly samples.

All datasets are red at low frequencies, although not sufficiently so to present problems of leakage from low to high frequencies in the time-series analysis programs we have used. Nonetheless, we have prewhitened all datasets by first-differencing.

Our spectral and cross-spectral analyses are based on the program of Carter and Ferrie (1979). The data are divided into n consecutive 50% overlapped blocks, to each of which a cosine "bell" has been applied after removal of the mean. Spectral and cross-spectral estimates are then obtained by application of a fast Fourier transform to each block and averaging over the blocks. The resulting number ν of degrees of freedom is given by $\nu = 18n/11$ (Welch, 1967).

In the present problem we took 13 overlapping blocks of 512 points, covering the period from 0130 LST 1 January 1982 to 0830 LST 29 May 1982. This gave 21 degrees of freedom.

The spectral and cross-spectral estimates are then used in the determination of the complex frequency-dependent coefficients a, b, c in a multiple regression,

$$\zeta = aP + bE + cN + \text{noise incoherent with } P, E, N, \quad (3.1)$$

of sea level on the three meteorological inputs. Doing this involves inversion of the Hermitian cross-spectral matrix between variables P, E and N (e.g., Wunsch, 1972; Garrett and Toulany, 1982; Palumbo and Mazzarella, 1982). Confidence limits on the computed values for the amplitudes of a, b, c are determined from the formula proposed by Garrett and Toulany (1982).

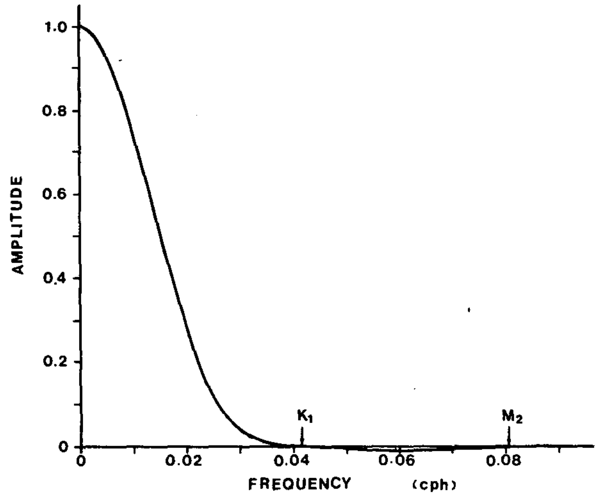

FIG. 4. Amplitude response of the $A_{24}{}^2A_{25}/(24^2 \times 25)$ tide-killing filter (see Godin, 1972).

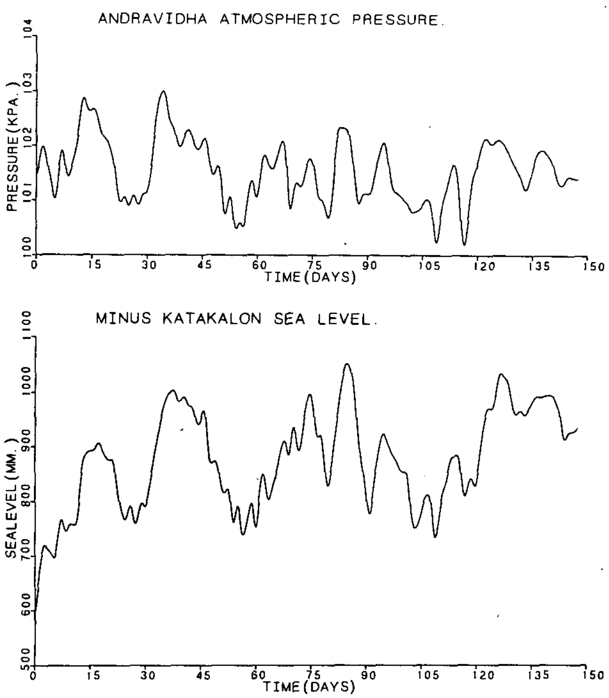

FIG. 3. (a) Filtered atmospheric pressure data from Andravidha; (b) minus the filtered sea level data from Katakolon. The start time is 1200 LST 2 January 1982.

b. Results

The power spectra of the prewhitened atmospheric pressure and sea level are shown in Fig. 5; the power spectra of the raw data may be obtained, to a very good approximation, by multiplying the spectra of the prewhitened data by ω^{-2}, with ω the frequency in radians h^{-1}. Fig. 6 shows the coherence (corrected for bias according to the formula of Jenkins and Watts, 1968), phase and gain. The gain and phase, defined by $\Phi_{P\zeta}(\omega)/\Phi_{PP}(\omega)$, with $\Phi_{ij}(\omega)$ the cross-spectrum between variables i and j, give the complex regression coefficient of sea level ζ on atmospheric pressure P that arises if other inputs such as wind are ignored. We have omitted error bars in Fig. 6; the results are shown mainly for comparison with the gain and phase,

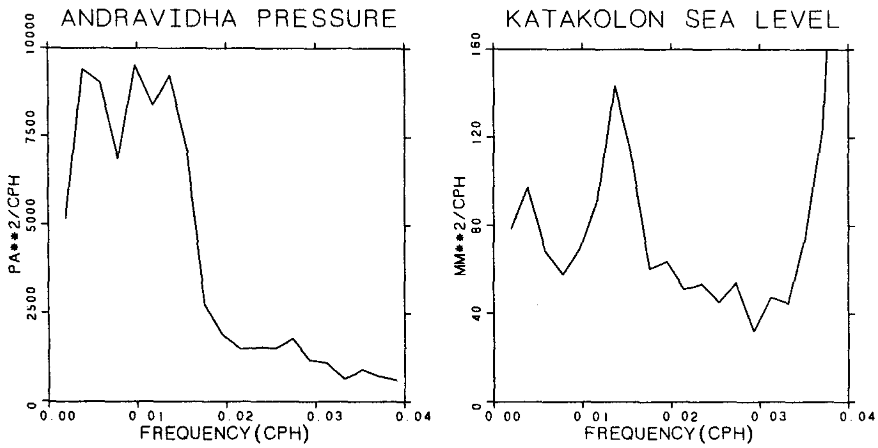

FIG. 5. Power spectra of prewhitened (a) Andravidha atmospheric pressure and (b) Katakolon sea level. The 95% confidence factors, for 21 degrees of freedom, are (0.59, 2.0).

shown in Fig. 7, that are determined if wind effects are allowed.

We do not show the other cross-spectra that are used in determining the multiple regression coefficients. However, it is worth remarking that while P is incoherent with E (except for some coherence above about 0.025 cycles per hour), it does have a significant coherence (about 0.6) with N, as is to be expected for

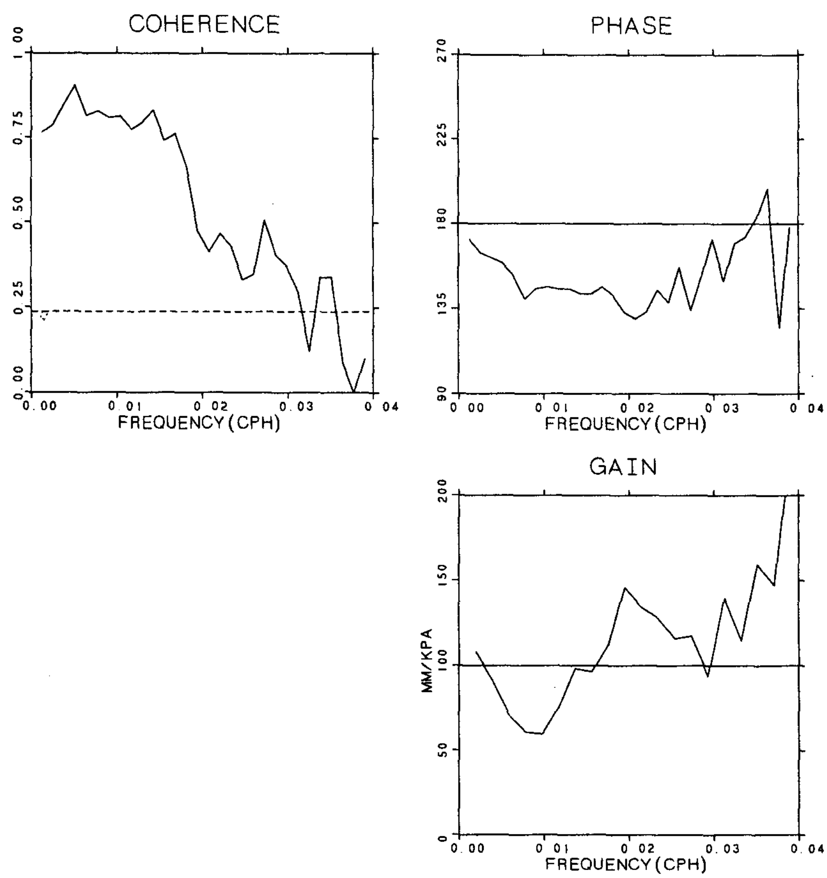

FIG. 6. Coherence, phase, in degrees, and gain for Katakolon sea level and Andravidha atmospheric pressure. The dashed line indicates the level above which the coherence is significant at the 95% confidence level. A Phase less than 180° corresponds to a lag of minus sea level behind atmospheric pressure.

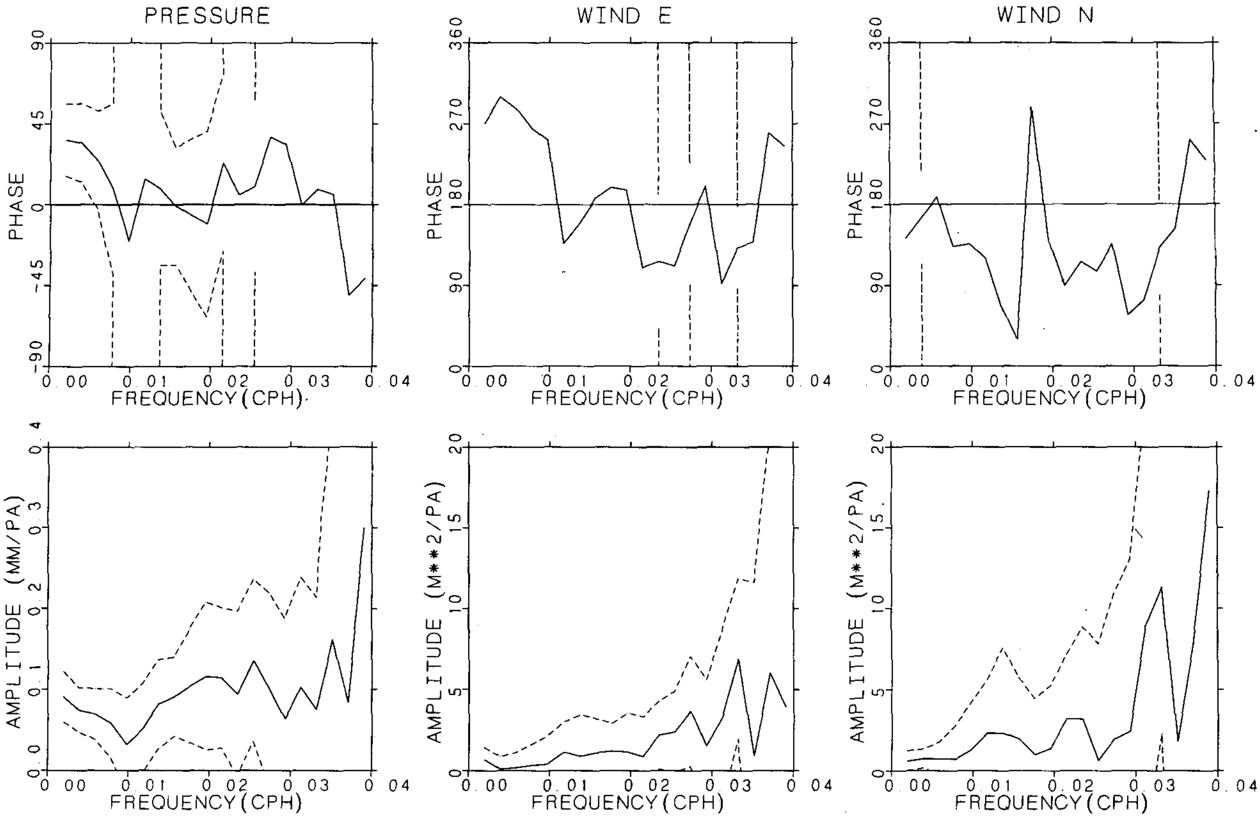

FIG. 7. Amplitude and phase of the regression coefficients of minus sea level on local atmospheric pressure P, wind E towards the East and wind N towards the North. Positive phase indicates a lag of minus sea level behind the input. The dashed lines are the 95% confidence limits which are not determined for the phase if the confidence limits on the amplitude do not exclude zero.

eastward propagating disturbances. Hence any sea level response to a northward geostrophic wind would tend to affect the gain in a simple regression of ζ on P alone. The objective in the use of multiple regression is to remove this effect.

The results of the multiple regression are shown in Fig. 7. The 95% confidence limits are rather distressingly large, but we do note that the sea level response to atmospheric pressure has a significant lag at low frequency. The amplitude of the pressure coefficient is only significantly different from the isostatic value of 0.1 mm Pa^{-1} (1 cm mb^{-1}) at 0.01 cycles per hour with the frequency resolution of our computations, but as the estimate at each frequency is independent, we conclude that the trend shown in the pressure coefficient, from about 0.1 at very low frequency, to a number less than this at 0.01 cycles per hour, is real. Above ~0.015 cycles per hour the coefficient is not distinguishable from 0.1.

A comparison of the pressure coefficient and phase from Fig. 7 with those of Fig. 6 shows that allowing for the effect of wind has led to a slight decrease in the amplitude of the pressure coefficient, but it still shows a phase lag at low frequency where it is significant.

The multiple regression coefficients for the dependence of sea level on wind are shown by the results displayed in Fig. 7 to be very poorly determined. Indeed they are not significantly different from zero. The comparative unimportance of wind, for this particular location, is supported by an analysis of the amount of sea level variance that can be attributed to the various inputs. If we multiply (3.1) by its complex conjugate, on both sides, we obtain

$$\begin{aligned}\zeta\zeta^* &= aa^*PP^* + bb^*EE^* + cc^*NN^* \\ &\quad + (bc^*EN^* + b^*cE^*N) + (ca^*NP^* \\ &\quad + c^*aN^*P) + (ab^*PE^* + a^*bP^*E) \\ &\qquad + \text{residual variance}, \qquad (3.2)\end{aligned}$$

where PP^* is to be taken as the power spectrum of P, and similarly for the other terms. The cross terms are related to the cross-spectra of the input variables. Dividing both sides of (3.2) by $\zeta\zeta^*$ and multiplying by 100 leads to an estimate of the percentage contribution to sea level of the various terms, as a function of frequency (Fig. 8). For frequencies less than 0.02 cycles per hour the residual variance averages less than 30% of the total variance, with most of the variance that

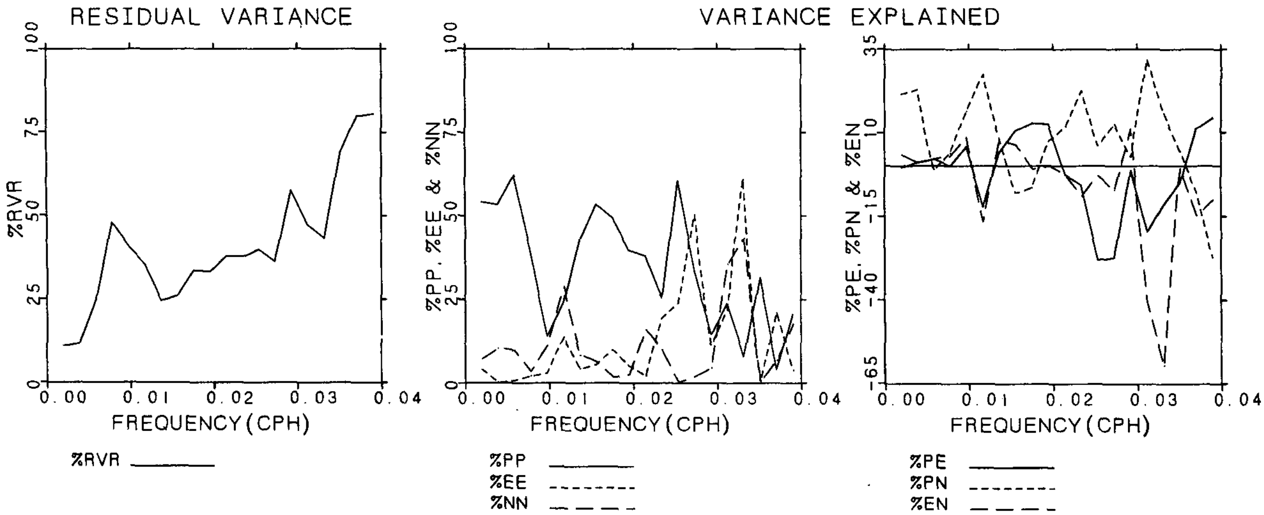

FIG. 8. Residual variance and the percentage of sea level accounted for by the various inputs. The contributions labeled P, E, N are solely from these inputs; EN, NP and PE from the combinations of inputs [see (3.2).]

is explained being attributable to the atmospheric pressure. At higher frequencies the residual variance increases and, as shown in Fig. 7, none of the regression coefficients is well determined.

4. Theory

In a theoretical study of the response of Mediterranean sea level to atmospheric pressure, Garrett (1983) used the simple model of Garrett and Toulany (1982) (see also Toulany and Garrett, 1984) for the volume flux $\mathrm{Re}[Qe^{-i\omega t}]$ through a strait of depth H, width W and length L in response to a sea level difference $\mathrm{Re}[\Delta\zeta e^{-i\omega t}]$. The general formula is

$$Q = gH\Delta\zeta[(-i\omega + \gamma)(L/W) + f]^{-1}, \quad (4.1)$$

where γ is the coefficient of linearized bottom friction. Appropriate values for H, L, W in this idealized model could probably be determined by running a numerical model for a particular strait. Garrett (1983) took $H = 250$ m for the Straits of Gibraltar and Sicily and $L/W = 3$. For reasonable estimates of γ the term $\gamma(L/W)$ is small compared with f. The term $-i\omega(L/W)$ is equal in magnitude to f at a period of 2.5 days if we take $f = 8.6 \times 10^{-5}\ \mathrm{s}^{-1}$ and $(L/W) = 3$. Thus for very low frequency motions (4.1) reduces to $Q = gH\Delta\zeta f^{-1}$, independent of L/W and ω. This approximate formula was used by Garrett (1983) for simplicity. In the present study we neglect γ but otherwise retain (4.1), although the inclusion of ω makes little difference.

The other main ingredient of Garrett's (1983) theory was the assumption that, within each basin of the Mediterranean, the sub-surface pressure, $P + \rho g\zeta$, with P the local atmospheric pressure and ζ the local sea level, is rendered uniform by rapidly propagating surface gravity waves, which would take only a few hours to traverse a basin. In fact the gravity waves involved will be Kelvin waves, so that while $P + \rho g\zeta$ may be made uniform around the edge of each basin, out to a distance of the order of the Rossby radius, it is possible for sea level in the center of a large basin not to experience changes in response to atmospheric pressure. For the Mediterranean, the Rossby radius is approximately 1500 km which is greater than the typical distance from a point in the interior of a basin to the nearest shoreline. Thus, to a fairly good approximation, we may assume that the whole of each basin is involved in a barotropic response to atmospheric pressure. If flow through the Straits is given by (4.1), then the sea level response of the Eastern and Western basins, of areas A_1 and A_2, respectively, depends particularly on the two parameters $\epsilon_i = \omega f A_i (gH_i)^{-1}$, with H_1, H_2 the depths at the Straits of Gibraltar and Sicily.

The other key parameter in Garrett's (1983) theory is the speed c with which atmospheric pressure systems move eastward across the Mediterranean. This enters the dimensionless parameters $\delta_i = \frac{1}{2}(\omega/c)L_i$, where L_1, L_2 are the lengths of the Eastern and Western basins, separated at the Strait of Sicily. (Our idealized model thus treats the Mediterranean as being made up of two rectangular basins of lengths L_1 and L_2, as shown in Fig. 9).

A minor extension of Garrett's (1983) theory, to allow for the ω-term in (4.1), leads to the following

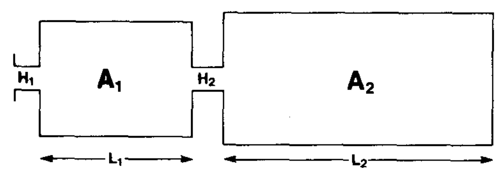

FIG. 9. An idealized representation of the Mediterranean. The western and eastern basins have areas A_1 and A_2, lengths L_1 and L_2, are separated from the Atlantic by the Strait of Gibraltar of depth H_1, and from each other by the Strait of Sicily of depth H_2.

expression for the response of sea level ζ to local atmospheric pressure P at a point, corresponding to Greece, halfway along the Eastern basin:

$$-\rho g \frac{\zeta}{P} = 1 - \delta_2^{-1} \sin\delta_2 + (1 - i\epsilon'_1)F\delta_2^{-1} \sin\delta_2 + i\epsilon'_1 F \exp[-i(\delta_1 + \delta_2)]\delta_1^{-1} \sin\delta_1, \quad (4.2)$$

where

$$F = [1 - \epsilon'_1\epsilon'_2 - i\epsilon'_1(1 + R) - i\epsilon'_2]^{-1} \quad (4.3)$$

$$\epsilon'_i = \epsilon_i\left[1 - i\left(\frac{\omega}{f}\right)\left(\frac{L}{W}\right)\right] = \epsilon_i\left[1 - i\epsilon_1 gH_1 f^{-2}A_1^{-1}\left(\frac{L}{W}\right)\right], \quad (4.4)$$

assuming the same value of L/W for each strait. The first two terms in (4.2) give the local response to atmospheric pressure in the absence of flow through the Strait of Sicily; this response is zero at low frequency ($\delta_2 \to 0$) and tends to 1 at high frequency ($\delta_2 \to \infty$) for which the scale of the atmospheric pressure pattern is much less than the size of the Eastern basin.

The third and fourth terms in (4.2) are a consequence of flow through the Straits of Gibraltar and Sicily. The fourth term disappears if we assume that the local atmospheric pressure (in Greece) is incoherent with the average atmospheric pressure over the Western basin. For a frequency tending to zero, the third term tends to 1 and the fourth term to zero, whereas at high frequency both terms tend to zero with F.

The formula (4.2) thus shows an isostatic response at low and high frequencies, with a nonisostatic response for intermediate frequencies.

Some of the parameters occurring in (4.2)–(4.4) are easily set. We take $A_2/A_1 = 2.0$ and $\delta_2/\delta_1 = L_2/L_1 = 2.0$. For values of L/W in a reasonable range the response calculated from (4.2) is only sensitive at high frequencies to the precise value of L/W. We take $L/W = 3$, so that with $H_1 = 250$ m and $A_1 = 8.4 \times 10^{11}$ m^2, $gH_1 f^{-2}A_1^{-1}(L/W) = 1.18$ in (4.4). The response in (4.2), for $-\rho g\zeta/P$ as a function of ϵ_1, is not very sensitive to values of ϵ_2/ϵ_1 in the range of 1 to 3. This is basically because, for an isostatic response of the Mediterranean, more flux is required through the Strait of Gibraltar than through the Strait of Sicily, so that the former is the more significant control. The results shown in Fig. 10 are for $\epsilon_2/\epsilon_1 = 2.0$, corresponding to $H_1 = H_2$.

The computed values of $-\rho g\zeta/P$ shown in Fig. 10 are for various choices of $\delta_1/\epsilon_1 = gH_1L_1(2cfA_1)^{-1}$, to which the results are clearly rather sensitive. If we take $L_1 = 1000$ km and $H_1 = 250$ m the values $\delta_1/\epsilon_1 = 0$, ½, 1, 2, correspond to $c = \infty$, 34, 17, 8 m s^{-1}.

The phase of $-\rho g\zeta/P$ in Fig. 10 is not particularly sensitive to the inclusion or neglect of the fourth term in (4.2), but the minimum amplitude is slightly less if the fourth term is ignored; i.e., if the local atmospheric pressure in the eastern basin is uncorrelated with the average pressure over the western basin. In that case the factor $\delta_2^{-1} \sin\delta_2$ refers not just to propagation of the pressure patterns, but to any spatial variability that reduces the average pressure over the eastern basin to a factor $\delta_2^{-1} \sin\delta_2$ of the local pressure.

5. Comparison of observations and theory

a. Pressure

In spite of the imprecision of the results from analysis of the sea level data and the crudeness of the theory, it is clear that there is some qualitative agreement between the observed response of sea level to pressure (Fig. 7) and the theoretical prediction (Fig. 10). A more precise comparison depends on choosing an appropriate value for δ_1/ϵ_1, or equivalently, for the speed c of eastward propagation of the atmospheric pressure patterns as well as determining the correlation between local pressure and average western-basin pressure.

Papa (1978) cites typical propagation speeds of 4 to 9 m s^{-1} for the western Mediterranean in summer, 11 to 18 m s^{-1} in winter. The NAVAIR Atlas (Anon, 1974) indicates a typical speed of 10 m s^{-1} for cyclonic disturbances in the January–May period of interest, but with a direction that is scattered about westerly so that the eastward trace speed will be somewhat larger than 10 m s^{-1}.

We have performed cross-spectral analyses between the time series for atmospheric pressure at the three Greek stations used in this study. The phases show a linear trend with frequency over the frequency range for which the signals are coherent (up to ~0.02 cycles per hour), the slopes of the trends indicating that Samos lags Andravidha by 4.2 h and Andravidha lags Kerkira by 2.9 h with Samos lagging Kerkira by 7.1 h consistently. These lags can be fitted by a wave propagating towards 142 deg (i.e., southeast) at 21 m s^{-1}. The eastward trace speed is thus 34 m s^{-1}!

Perhaps more relevantly, we obtained atmospheric pressure data for Cagliari, Sardinia, and computed the cross-spectrum between it and Andravidha pressure (Fig. 11). There is some coherence for frequencies less than about 0.005 cycles per hour, with a phase lag that increases more or less linearly with frequency, in spite of low coherence, and corresponds to a time lag of about 15 h, or a propagation speed of about 20 m s^{-1}.

In summary, the curves in Fig. 10b [neglecting the fourth term in (4.2)] are somewhat more relevant than those in Fig. 10a, though for very low frequencies some attention should be paid to Fig. 10a with $\delta_1/\epsilon_1 \simeq 0.9$ (for $c = 20$ m s^{-1}). In the use of Fig. 10b it is appropriate to think of δ_1/ϵ_1 as representing the scale of the pressure patterns, independently of their propagation speed or spatial coherence. The best correspondence between the theoretical predictions of Fig. 10b and the observed phase and amplitude of the pressure coefficient in Fig. 7 is for a value of δ_1/ϵ_1 between 1 and 2, although the

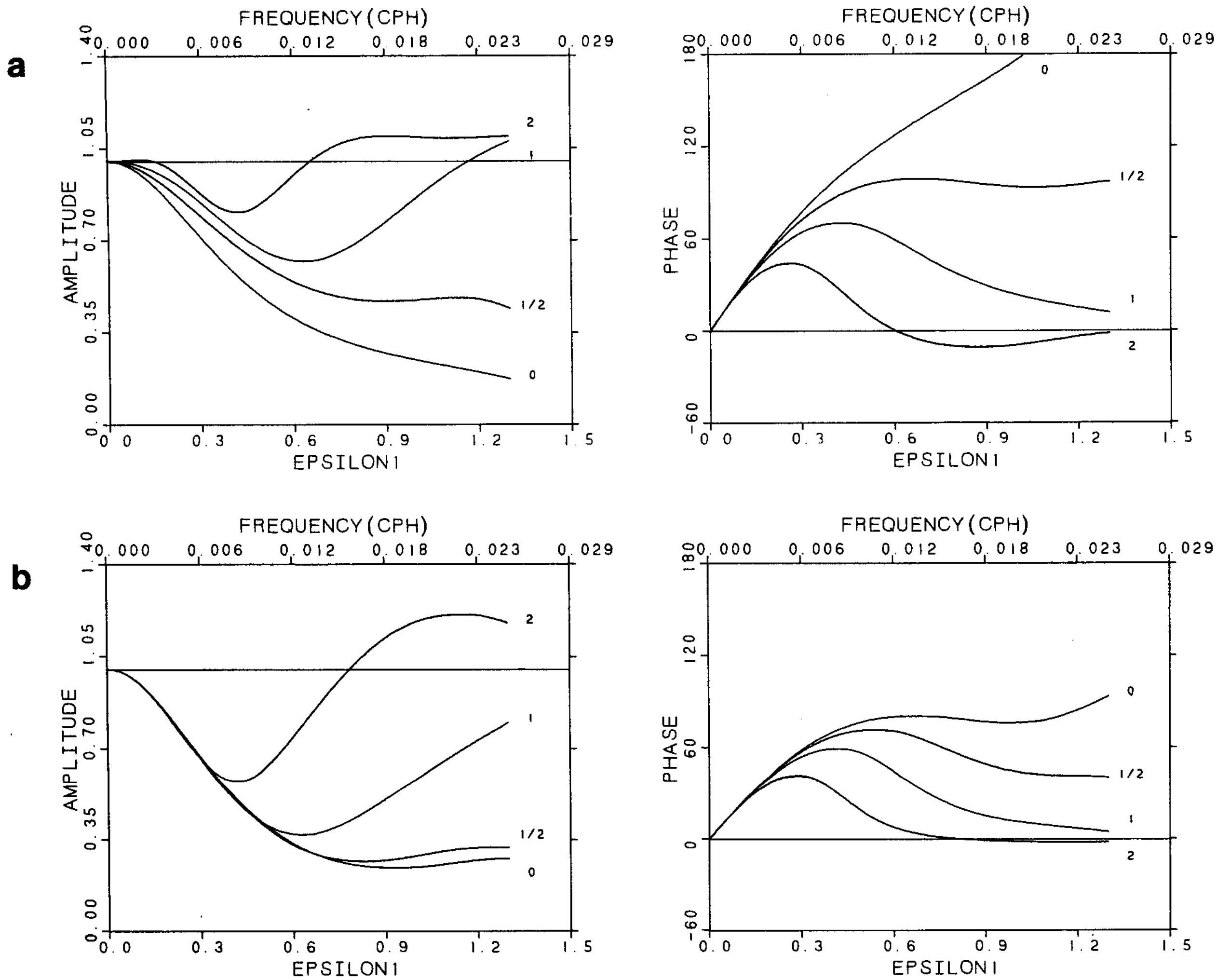

FIG. 10. Prediction of $-\rho g\zeta/P$ for a position, representing Katakolon, halfway along the eastern basin in the model shown in Fig. 9. The values are determined from (4.2) with $A_2/A_1 = 2.0$, $\delta_2/\delta_1 = 2.0$, $\epsilon_2/\epsilon_1 = 2.0$ and $gH_1f^{-2}A_1^{-1}(L/W) = 1.18$ in (4.4). The conversion of ϵ_1 to frequency assumes $H_1 = 250$ m, for which $\delta_1/\epsilon_1 = 0$, ½, 1, 2 (marked) corresponds to $c = \infty$, 34, 17, 8 m s^{-1}. (a) The fourth term in (4.2) is included, whereas in (b) it is omitted as for zero coherence between local pressure and the average pressure over the western basin.

error bars in the observed values and the simplified nature of the theory make a precise matching unjustified. A value of 1.5 for δ_1/ϵ_1 corresponds to $c = 11$ m s^{-1} and a wavelength of 4000 km at 0.01 cycles per hour, which is not inappropriate.

The agreement between theory and observation, within the limitations of both, is quite good. Both show an amplitude of the normalized pressure coefficient that is unity at very low frequency, falling to ½ or less at about 0.01 cycles per hour before rising to unity again. The lag of sea level behind minus atmospheric pressure is 0° at very low frequency (though this is not resolved by the data), rising to a few tens of degrees then falling back to zero by about 0.01 cycles per hour.

b. Wind

We remarked in Section 3 that the regression coefficients of Katakolon sea level on geostrophic wind are not significantly different from zero (Fig. 7) and that the winds do not account for much of the sea level variance. Nonetheless, we notice that the phases of the wind coefficients at low frequency in Fig. 7 are best taken to be 180°, which corresponds to a raising of sea level by northward and eastward geostrophic winds; i.e., by surface winds which are onshore or longshore with the coast to the right. The gain is larger for the longshore wind component. Both of these results are qualitatively consistent with theory (e.g., Csanady, 1982).

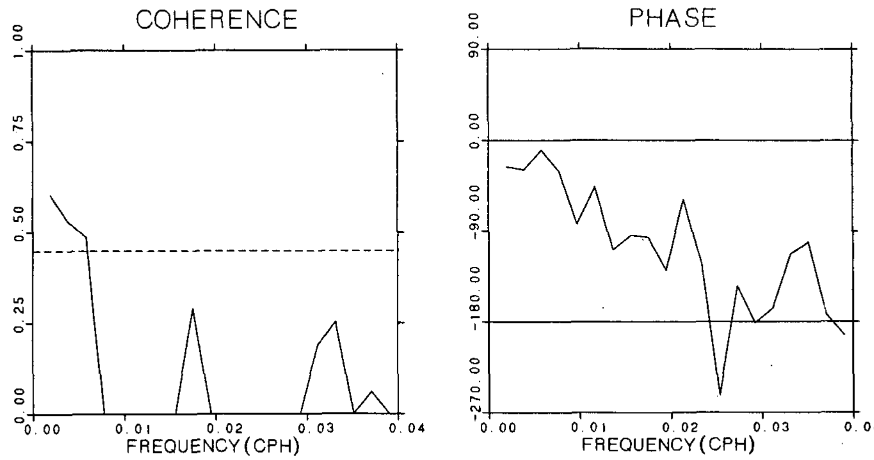

FIG. 11. Coherence and phase between the time series for atmospheric pressure at Cagliari, Sardinia and Andravidha, Greece. Negative phase corresponds to Andravidha lagging Cagliari.

Quantitatively, we may take the response to the northward geostrophic (i.e., longshore surface) wind to be ~2 m^2 Pa^{-1}, i.e., 2×10^{-4} m for a geostrophic wind of 1 m s^{-1} or a surface wind of about 0.7 m s^{-1}. This would correspond to Sandstrom's (1980) formula for the coastal set-up of a frictionally balanced longshore current for a shelf-width of only 1 km. The seafloor slope offshore from Katakolon and its vicinity is indeed very steep, with a depth of 200 m being achieved within a few kilometers (Fig. 2).

6. Discussion

Analysis of sea level and meteorological data from Katakolon, Greece, has shown a significantly non-isostatic response to atmospheric pressure at a frequency of about 0.01 cycles per hour, although an isostatic response is achieved at very low, or much higher, frequencies. The results are consistent with the predictions of a simple theory in which fluctuating flow through the Straits of Gibraltar and Sicily is parameterized in a simple manner that corresponds to "geostrophic control" at low frequency.

More elaborate data analyses could be performed if one were to assemble larger datasets on sea level and atmospheric pressure. In particular, it would be interesting to use as an input to the multiple regression the sea level outside the Straits of Gibraltar. The present theory assumes this to be purely isostatic; any wind-driven departure from isostasy (which may be a moderate fraction of the response to pressure, see Garrett, 1983), and its effect on the Mediterranean, is probably a contributor to the residual variance at Katakolon. (It does not seem to have emerged via local winds or pressure, which are almost certainly incoherent with winds outside the Straits of Gibraltar.)

The present multiple regression technique runs the risk of assigning a response to one of the inputs that is actually due to some other correlated variable that has not been considered. At low frequency this could be a change in mean sea level of the Mediterranean associated with changes (particularly seasonal) in the difference between evaporation and precipitation. The frequency at which this might occur is probably lower than the lowest (0.002 cycles per hour) resolved in the present study, although the results of Palumbo and Mazzarella (1982) suggest that at lower frequencies there are such changes in sea level. It would be interesting to examine in detail the seasonal changes in sea-level difference between Atlantic and Mediterranean and relate them to changes in flow through the Strait of Gibraltar and the annual cycle of other oceanographic and meteorological variables.

Acknowledgments. We are most grateful to Admiral M. Poulidis, Hydrographer of the Hellenic Navy, for providing the Greek data used in this study. We also thank David Phillips of the Canadian Climate Centre for providing the Cagliari pressure data. Bechara Toulany provided access to computer programs and considerable assistance. Our work is supported by the Natural Sciences and Engineering Research Council of Canada.

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