# Small-Scale Variations of the Wind-Driven Coastal Sea-Level Response in the West Florida Bight

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## ABSTRACT

Records from a number of temporary tide gage stations spaced an average of 20 km apart near Cedar Key, Florida are used to examine the alongshore pressure gradient over length scales much smaller than before possible. In agreement with previous studies, an alongshore stress of 1 dyn $\mathrm{cm}^{-2}$ will produce a large-scale (over distances of several hundred kilometers) alongshore sea-level gradient on the order of $10^{-6}$, sea level rising downwind. However, cross-shore stress is also important and, at short periods (~3.5 days), can be the more coherent forcing. Estimates of small-scale sea-level slope do appear to show systematic deviations from the larger scale slope, some of which seem to be related to alongshore variations in nearshore bathymetry.

## 1. Introduction

Recent studies of continental-shelf circulation have shown that persistent longshore pressure gradients along open coasts play an important dynamical role (e.g., Csanady, 1981). Theoretically, under particular constant distributions of wind stress, sea levels adjust in a coastal boundary layer giving rise to significant cross-shore and alongshore pressure gradients. Evidence for these gradients is often sought by examining alongshore differences between suitably smoothed tide gage records. Typically, the stations are several hundred kilometers apart and the differences are on the order of a few centimeters, so that slopes of about $(1-2) \times 10^{-7}$ are obtained. If the signal is large enough, differences over smaller alongshore distances ought to be measurable as well; in such cases, a natural question to ask is whether similar values of sea level slope will be found or whether meaningful small-scale variations will be present.

Over the very broad West Florida Shelf (Fig. 1), the winter atmospheric forcing is particularly distinct and repetitive, giving rise to a clear sea-level and current response as shown in recent observational studies (Cragg *et al.*, 1982; Marmorino, 1982; Mitchum and Sturges, 1982). Analyses of widely spaced tide-gage data suggest that relatively large alongshore sea-level slopes, in the range $(5-10) \times 10^{-7}$, arise in response to episodes of alongshore wind stress, typical coastal values being less than 1 dyn $\mathrm{cm}^{-2}$. Recent numerical model studies (Hsueh *et al.*, 1982; Marmorino, 1982) nicely reproduce these large-scale gradients; in addition, they show small-scale reversals in the slope, apparently as a result of alongshore variations in the nearshore bathymetry with length scales on the order of 20 km (comparable to the model resolution). These variations are particularly distinct in the area that I call the "West Florida Bight": that part of the shelf between the Tampa Bay/St. Petersburg area to the south and the "Big Bend" area (where Apalachicola and Shell Pt. are located) to the north (Fig. 1). In this paper, records from a set of closely spaced gages located in the Bight are examined to see if such data may be at all useful. An equivalent experiment away from shore would be to measure along-isobath sea-level slopes with bottom pressure gages (e.g., Hayes, 1979). Since coastal data are already in hand, it seems reasonable first to examine them.

The paper is organized as follows: Section 2 presents the available data sets; the tidal content of the sea-level data is examined in Section 3 with the purpose of insuring that the data are of good quality; Section 4 deals with the low-frequency sea-level fluctuations and their relation to the wind-forcing during two consecutive study periods that differ in both the strength and nature of the forcing; and Section 5 deals explicitly with the calculation of alongshore slope. Section 6 compares the new results with previous observations and model calculations.

## 2. Data

As part of a cooperative effort between the National Ocean Survey and the Florida Department of Natural Resources to determine precisely mean high and low water lines, hundreds of closely-spaced temporary tide-gage stations have been occupied along the coast of Florida. The ones chosen for this exploratory anal-

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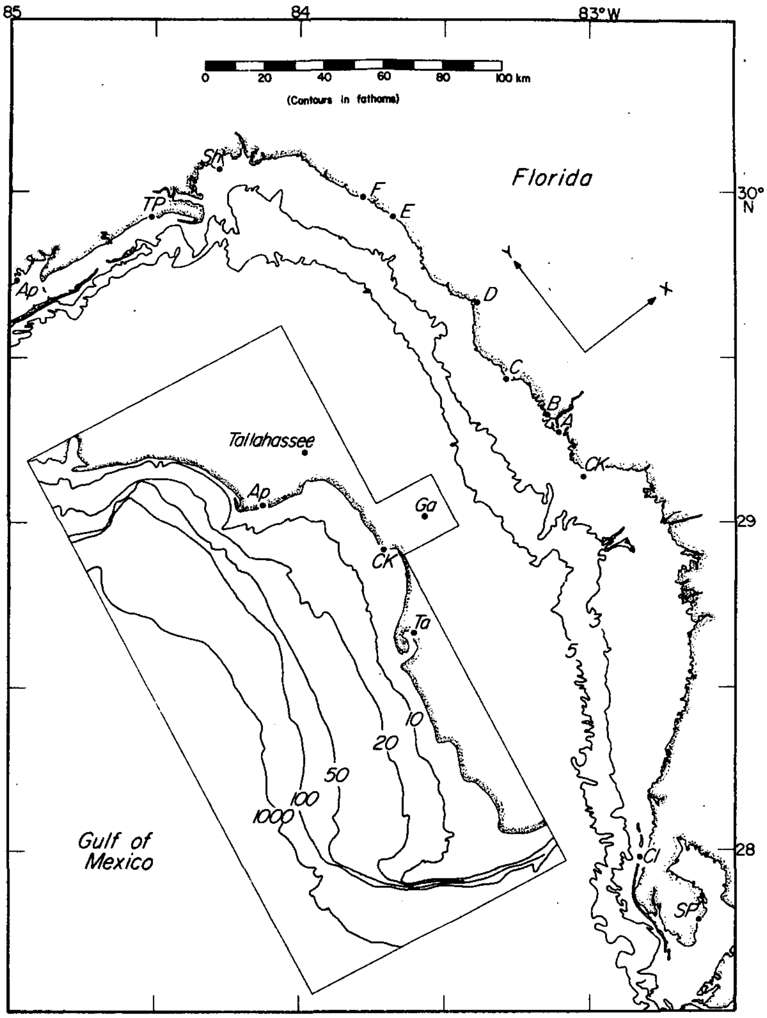

FIG. 1. Location of tide gages in the West Florida Shelf Bight and coordinate system used for analysis. The $y$ axis, aligned with the local coastline, points toward 323°T. Inset shows the larger scale bathymetry (contours in fathoms) and the meteorological stations in the area. Abbreviations are: Ap (Apalachicola), TP (Turkey Point—Florida State University Marine Lab), Sh (Shell Point), F (Fenholloway River), E (Spring Warrior Creek), D (Steinhatchee), C (Horseshoe Point), B (Suwannee River), A (Suwannee River entrance), CK (Cedar Key), Cl (Clearwater), SP (St. Petersburg), Ga (Gainesville) and Ta (Tampa).

ysis lay along a fairly straight stretch of coast, near the permanent control station at Cedar Key (Fig. 1). Recent current measurements (Mitchum and Sturges, 1982) offshore of this coastline make this set of data of particular interest. Eleven stations were occupied in this area from June through December 1978; however, equipment malfunctions and other problems limit the useful data to six stations, A–F (Fig. 1), for one common period, 1 November to 10 December 1978. Two of these stations, A and D, were left operating into March; thus, a second study period is available to compare with the first. In order to explore the sea-level response on a larger scale, records from tide stations to the north (at Shell Point), to the south (at St. Petersburg), and at Cedar Key were also analyzed. (The coastal station at Clearwater Beach, which has a more exposed location than the St. Petersburg gage, ceased operating after 30 November, and was thus not used.) Except for the tidal analysis of Section 3, this paper uses data that have been low-pass filtered to remove tidal and other high-frequency fluctuations and decimated to 6 h intervals. (Filter amplitude response is 95% at 66 h, 50% at 40 h, and 5% at 29 h.) As our primary interest is in sea-level slope among the closely spaced stations, no adjustment for atmospheric pressure was made. The effect of such an adjustment would be to lower the presented amplitude and gain values of Section 4 by ~15% (Marmorino, 1982). After filtering, the dates of the two study periods become 3 November–6 De-

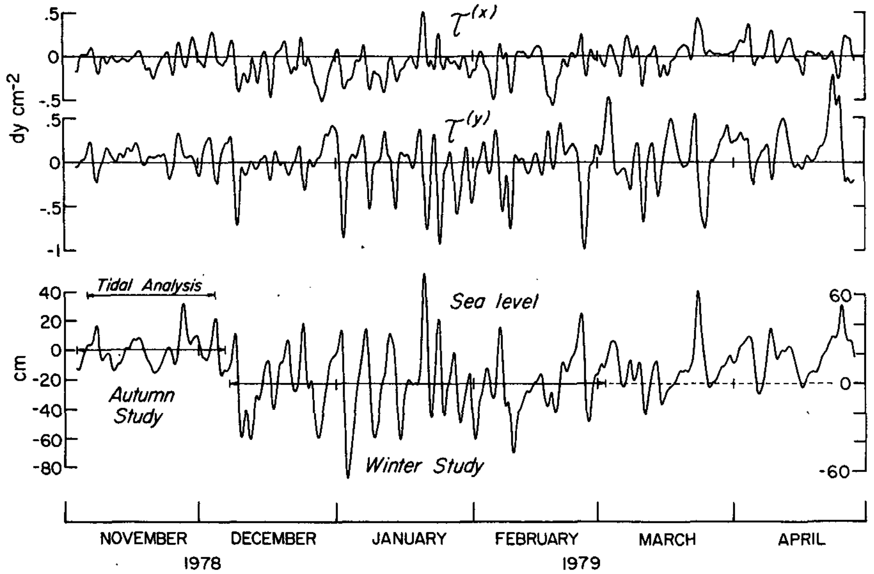

FIG. 2. Low-passed Apalachicola wind-stress components and Cedar Key sea level, with analysis periods indicated. Horizontal lines indicate mean values for the autumn and winter study periods (note the shift in the vertical scales). Note the lowering of mean sea level brought about by the stronger wind events of early December. Mean wind stresses over the winter period are −0.10 and −0.04 for the $x$ and $y$ components, respectively.

cember 1978 (33.25 days) and 8 December–2 March (84.75 days). These two studies span the 1978–79 late autumn-winter season as shown in Fig. 2, where the sea-level record at Cedar Key is presented. Records at the other stations are strikingly similar with 96% of the variance being accounted for by a single empirical mode (from orthogonal function analysis).

Meteorological data are available from five stations around the Bight (see inset in Fig. 1). The Cedar Key station is part of the remote DARDC network; unfortunately, its record had too many gaps to be useful. The other records are highly correlated, as expected from previous studies (Cragg *et al.*, 1982; Marmorino, 1982). The coastal station at Apalachicola was chosen as providing the closest approximation to overwater winds; indeed, the winds there were the most energetic. The 3 h data were converted to stress components in the $x$ and $y$ directions, approximately onshore and alongshore (Fig. 1), and low-pass filtered. A constant drag coefficient of 0.0014 was used in the stress computation. The stress time series are compared with the Cedar Key sea levels in Fig. 2. There is evidence that actual overwater stress values are as much as a factor of 2 or 3 greater than the coastal winds imply (Marmorino, 1982; Mitchum and Sturges, 1982). This must be taken into account later when values of sea-level response per unit stress are presented.

## 3. Tidal analysis

Prior to examining the wind-driven signals, an analysis of the tidal signals is made as a check on the accuracy of the data. A 29-day record length, centered within the autumn study period, was chosen to provide good spectral resolution of the $M_2$, $S_2$, $K_1$ and $O_1$ constituents. The amplitudes and phases at the three permanent tide-gage stations are consistent with previous determinations (Tables 1 and 2). Amplitudes at temporary stations C–F vary reasonably (a "noise" level of, say, ±1 cm) compared to values for Cedar Key and Shell Point (Table 1 and Fig. 3). Amplitudes at A and, to a lesser extent, at B are lower than the rest, plausibly on account of their location near the mouth of the Suwannee River, i.e., part of the signal propagates up-river and into the marshy coastal areas. Similarly, the $M_2$ and $S_2$ amplitudes at St. Petersburg, located in Tampa Bay, are ~40% less than at Clearwater Beach on the open coast (Fig. 1).

TABLE 1. Tidal amplitudes (cm) based on raw spectral estimates (no smoothing or data tapering) from a 29-day analysis beginning 0000 EST 6 November 1978. Values in parentheses have been determined by the National Ocean Survey using a 365-day analysis at Cedar Key and St. Petersburg and a 29-day analysis at Shell Point.

| Station | $M_2$ | $S_2$ | $O_1$ | $K_1$ |
|---|---|---|---|---|
| Sh | 36.9 (35.7) | 14.4 (12.0) | 12.4 (16.2) | 18.3 (16.8) |
| F | 37.9 | 14.8 | 12.1 | 18.6 |
| E | 36.9 | 14.2 | 11.4 | 17.8 |
| D | 37.6 | 15.7 | 13.2 | 18.7 |
| C | 36.2 | 14.3 | 13.1 | 18.0 |
| B | 34.8 | 12.6 | 11.1 | 16.2 |
| A | 31.0 | 10.0 | 9.4 | 14.1 |
| CK | 38.3 (36.6) | 14.0 (12.9) | 13.5 (15.4) | 18.3 (16.8) |
| SP | 17.6 (16.1) | 5.2 (5.2) | 11.6 (14.4) | 16.6 (16.0) |

TABLE 2. Tidal phases relative to Cedar Key. Positive value means that Cedar Key leads. 95% confidence intervals are shown. In parentheses are phases with respect to Cedar Key, obtained from National Ocean Survey analyses: ($K_1$, $O_1$) and ($M_2$, $S_2$).

| Station | $K_1$ and $O_1$ | $M_2$ and $S_2$ |
|---|---|---|
| Sh | 1 ± 5 (8, 8) | 15 ± 1 (25, 23) |
| F | −1 ± 3 | 8 ± 1 |
| E | 2 ± 3 | 9 ± 2 |
| D | −7 ± 2 | −2 ± 1 |
| C | −4 ± 2 | −4 ± 2 |
| B | 2 ± 3 | −3 ± 2 |
| A | 11 ± 3 | 7 ± 3 |
| CK | 0 (0, 0) | 0 (0, 0) |
| SP | 9 ± 4 (17, 12) | 8 ± 7 (12, 3) |

This kind of amplitude reduction should not be a factor for the low-frequency wind-driven fluctuations. Semi-daily phases are seen to increase alongshore northward from the area between Cedar Key and D. Again, the influence of the Suwannee River and Tampa Bay can be seen at A and St. Petersburg, this time as phase delays with respect to the nearest "solid coastline" stations, Cedar Key and Clearwater, respectively. (At Clearwater, the daily and semi-daily components lead Cedar Key by about 25° and 66°, respectively.) In summary, the data from stations A–F appear to be of good quality.

## 4. Wind-forced response

During the late fall and winter, coastal sea level fluctuates in response to atmospheric cold fronts which move generally southeastward (in the negative $y$ direction) through the area every 6 days or so. Preceding each cold front, winds are generally northward or northeastward behind a warm front; the winds then rotate clockwise, finally blowing out of the northwest behind the cold front. Inspection of the time series in Fig. 2 reveals these repetitive episodes of wind stress that, alternately, raise and lower coastal sea level. The first study period, which includes all six of the temporary gages, involves the preliminary, relatively weak frontal passages of late autumn; the second study period begins with the onset of the larger-amplitude signals characteristic of stronger winter forcing.

### *a. Autumn study (3 November–6 December 1978)*

The relation between wind forcing and sea-level response at a typical station (A) is examined for the autumn period in Fig. 4. Highest coherence is found for a wind stress oriented nearly cross-shore (0°) at periods of about 3.5 days, where both wind stress and sea level are most energetic. A wind stress of 1 dyn cm$^{-2}$ with such an orientation would give rise to a sea-level response of about 60 cm. (This must, however, be corrected for the effects of atmospheric pressure and, probably, realistically greater values of wind stress, as previously discussed.) As shown, the sea level response at A would lag the Apalachicola stress signal by about 60° or 0.6 days (for the 3.5-day signal). The lag with respect to the local wind stress is estimated to be slightly less, about 10 h. (This has been done by using half the time lag between Apalachicola and Gainesville calculated for this time period.)

The spatial variation of the dominant sea level signal is shown in Fig. 5. The signal is highly coherent over the 300 km span of the stations. Propagation is southward, as expected, at about 850 km day$^{-1}$, so that the signal requires only 4 h to pass through the Bight stations. Within the Bight, there seems to be a systematic variation in the amplitude of the response, with a minimum occurring near D. Other ways of exploring this variation in amplitude give similar results. For example, Fig. 6 is based upon a calculation of variance in individual records; again, amplitude response is lower—by about 2 cm, or 13%—near D. The first mode from an empirical function analysis also shows the same kind of variation, as does a simple visual comparison between simultaneous events at different stations. Under the assumptions that the amplitudes have been correctly determined and that the response to a given wind event occurs nearly simultaneously at all the closely-spaced Bight stations, the instantaneous alongshore sea-level slope would vary about as shown in Figs. 4–6.

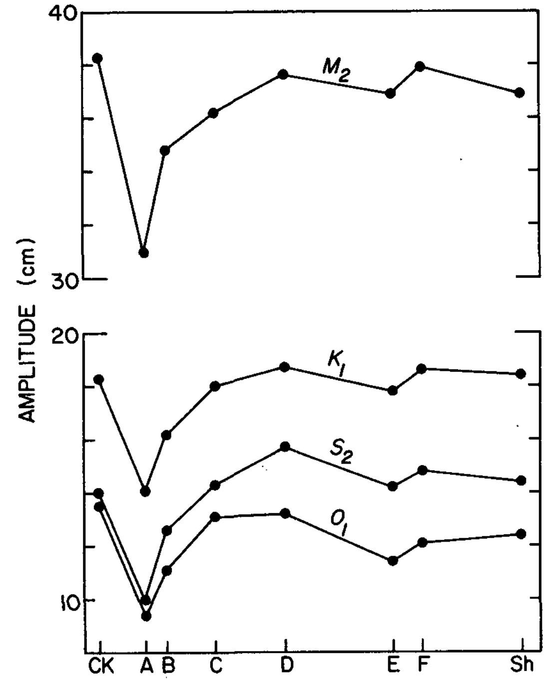

FIG. 3. Alongshore distribution of tidal amplitudes (see Table 1).

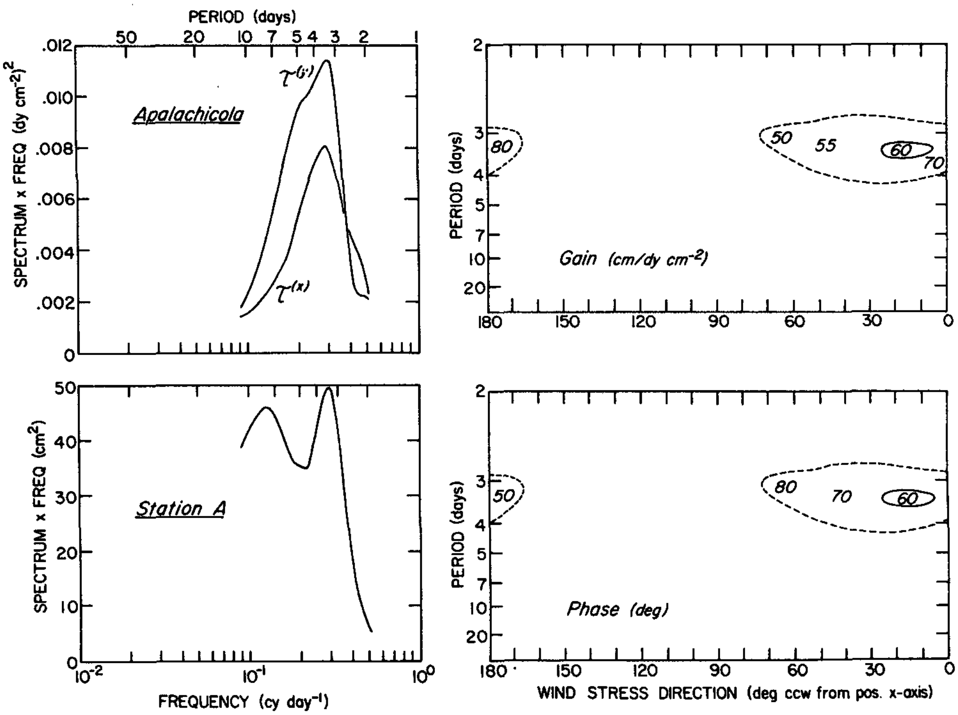

FIG. 4. Sea level at A vs Apalachicola wind stress for the autumn study period. Left: spectra of sea level and wind stress, plotted in equal-area, equal-variance form. Right: both panels show coherence squared between sea level and wind stress of various orientations (measured counterclockwise from the positive $x$ axis), contours being shown for values of 0.9 (solid curve) and 0.8 (dashed curve). Degrees of freedom $\approx 11$; thus, the 99% significance level for coherence squared is about 0.64. Note that the highest coherence is found for an orientation of about 15° or −165°. In the upper panel are values of gain (modulus of the transfer function) between sea level and wind stress of a particular orientation; in the lower panel are values of phase, which must be corrected before giving the lag between sea level at A and the local wind stress (see text). For coherence squared of 0.9, 95% confidence intervals for phase are ±14°; for coherence squared of 0.8, ±23°.

### *b. Winter study (8 December–2 March 1979)*

In response to stronger forcing, winter sea-level fluctuations are more energetic than during the autumn (Fig. 6), and while the response at Station D was smaller than at surrounding stations during the autumn, it is now larger than at Station A. A detailed examination of the A and D records shows (Fig. 7) a 5% greater response at D, in agreement with Fig. 6 and event-by-event comparisons. Signals propagate at about 825 km day$^{-1}$ from D toward A (about what was found in the autumn). The response appears to continue to increase in the positive $y$ direction, being somewhat larger at Shell Pt. than at D (Fig. 6). Note that the Cedar Key response is, on average, larger than at A.

In contradistinction to the autumn result, sea-level response is now more coherent with the alongshore wind (Fig. 8). This is clearest at periods > 5 days where the highest coherence is for alongshore (90°) orientations. At shorter periods (~3.5 days, again), coherence is high for a wider band of wind-stress orientations, including the cross-shore direction. At the long periods, sea level lags the Apalachicola stress by about 1.5 days, and the local stress by an estimated 1.3 days. At the shorter periods, sea level lags local alongshore stress by about 18 h, the local cross-shore stress by about 6 h. For a given orientation, gain increases as period increases, as was found by Marmorino (1982) in his 1978 winter study. For fixed frequency, gain increases as orientation shifts away from the alongshore direction, becoming largest for nearly cross-shore wind orientations. The largest gains found in the autumn study were for cross-shore orientations as well. Mitchum and Sturges (1982) also find that the gain between February–March 1978 Cedar Key sea level and an estimated local cross-shore stress is much higher [140 cm (dyn cm$^{-2}$)$^{-1}$] than with an alongshore stress [60 cm (dyn cm$^{-2}$)$^{-1}$].

## 5. Direct calculation of alongshore sea-level slope for the winter period

The evidence so far presented for wind-driven alongshore slope has been rather indirect, relying on the spatial variability of some statistical measure of sea-level *response*. Consider the case of the winter period for which the fluctuations are greater in num-

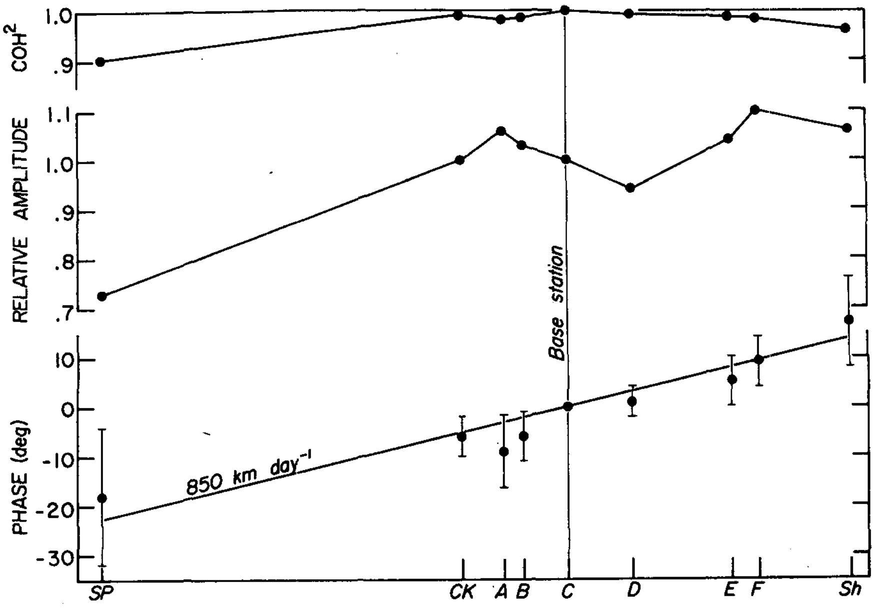

FIG. 5. Alongshore distribution of coherence squared, amplitude and phase lead for the 3.63-day sea-level signal that is dominant in the autumn study period. Values are measured relative to the middle "base" station C. Phase propagation is from Shell Pt. (on the right) toward St. Petersburg at about 850 km day$^{-1}$. Confidence intervals are 95% values. (Results are similar for the less important 8-day signal.)

ber and amplitude. Fig. 7 shows that the response at D is greater than at A (52 km distant). At the important synoptic frequencies, the response is mainly to alongshore stress. (This was shown in detail for A, but is just as true for D.) With this evidence one *expects* the following: when the wind stress is in the positive $y$ direction, shoreward Ekman transport will raise sea level more at D than at A so that sea level

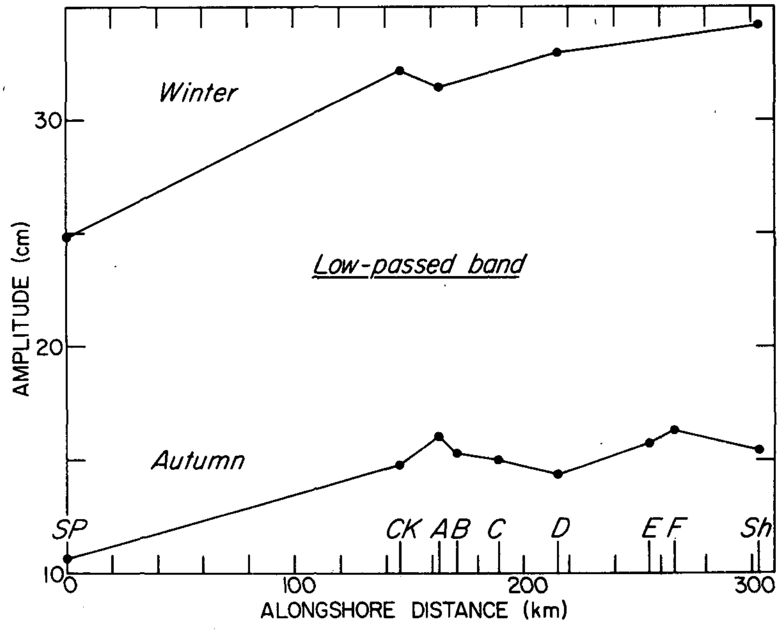

FIG. 6. Alongshore distribution of sea-level amplitude for the autumn and winter study periods. Amplitude = (2 × variance)$^{0.5}$, where the variance is based on the low-passed series. Distance is measured along the $y$ axis.

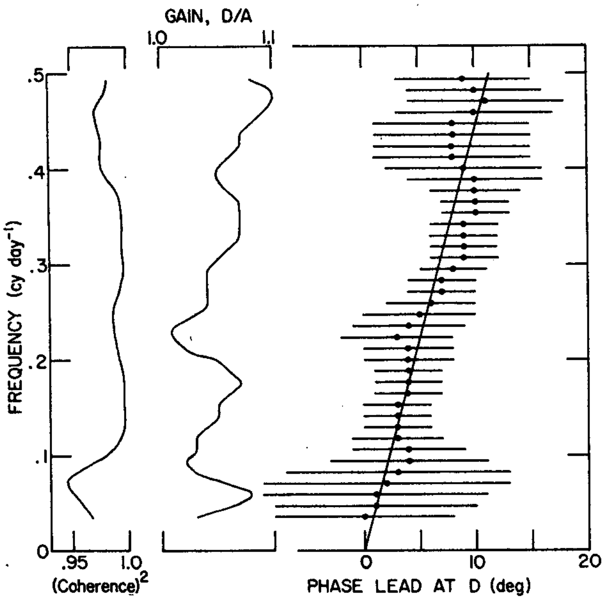

FIG. 7. Spectral analysis between sea level at Stations A and D for the winter study period. The slope of the straight line through the phase estimates (±95%) implies signal propagation toward A at about 825 km day$^{-1}$. Degrees of freedom ≈ 11.

will rise downwind, giving rise to a pressure force that opposes the wind stress; similarly, for wind stress in the negative $y$ direction, sea level will also rise downwind on account of the larger decrease at D. But will a comparison of the instantaneous slope between D and A (created by a simple differencing after adjusting both series to zero mean) to the alongshore stress show the expected behavior? The answer is (Fig. 9) that many of the events in sea-level slope are indeed correlated with alongshore wind stress in the expected way, but some are clearly not. On the other hand, slopes calculated over larger scales (158 and 303 km) are strongly correlated in the expected fashion with the alongshore stress (Fig. 9). Correlations between the small-scale slopes and the large-scale slope (taken to be the Sh–CK value) vary as follows: D–CK (69 km separation), −0.25; D–A (52 km), 0.54; and A–CK (17 km), −0.56. (A correlation coefficient greater than 0.50 is significantly different from 0 at the 1% level.) Table 3 shows that less than 44% of the variability of the small-scale slope signal is accounted for on the basis of a simple linear regression using both components of wind stress. For the A–CK and D–A slopes, the alongshore wind stress explains more of the variability; for the D–CK slope, the cross-shore wind stress is more important.

In summary, then, directly calculated small-scale slopes are variable and not easily explained in terms of wind forcing. The D–A slope is in the same sense as the large-scale alongshore-wind-driven slope, the A–CK slope is in the opposite sense, but both are in agreement with the response-type estimate of slope (e.g., Fig. 6); the D–CK slope is not significantly correlated with the large-scale slope. For small separa-

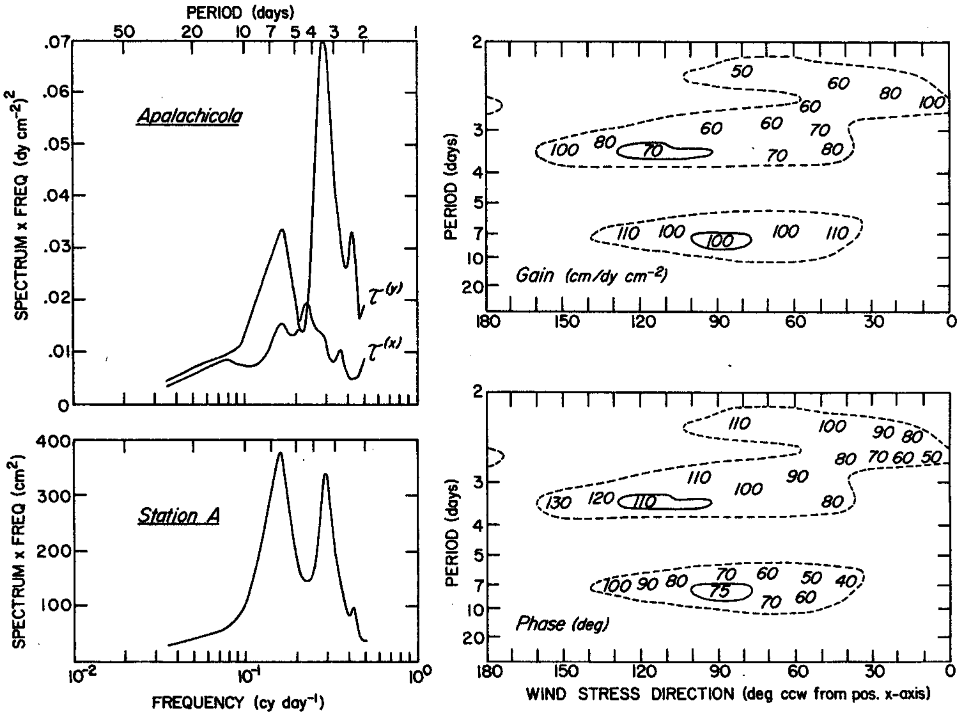

FIG. 8. Sea level at A vs Apalachicola wind stress for the winter study period. See caption for Fig. 4 for explanation.

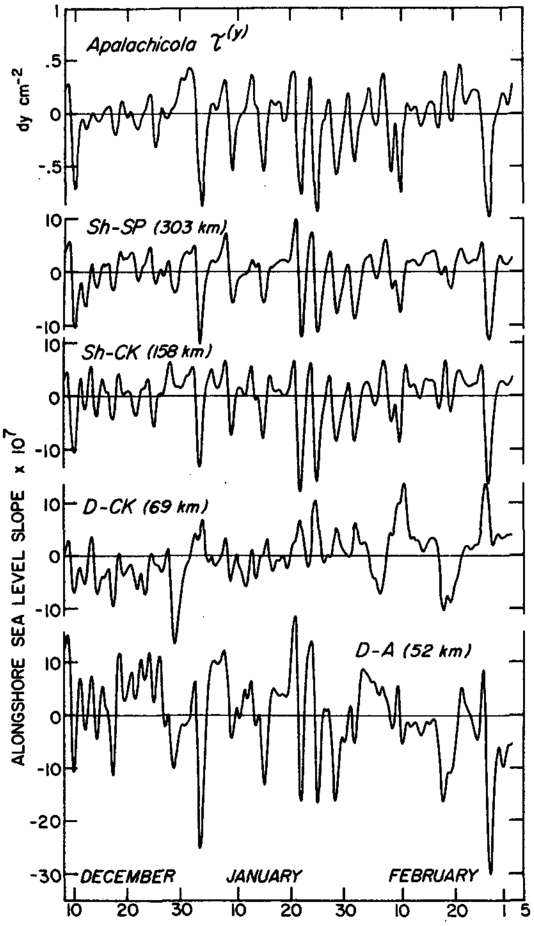

FIG. 9. Alongshore sea-level slopes during the winter study period in order of decreasing station-pair separation (values in parentheses). The plot for A–CK (17 km) is not shown. Abbreviations: Sh = Shell Pt, CK = Cedar Key, and SP = St. Petersburg. For comparison, the alongshore component of the Apalachicola wind stress is shown at top.

tions, slope calculated directly is based on relatively small sea-level differences; possibly, lower-frequency sea-level fluctuations that might otherwise be unimportant because of their small amplitude interfere with the wind-driven-slope signal, introducing noise that is comparable to the signal of interest. Notice that the small-scale slope series in Fig. 9 do appear to contain a low-frequency signal not present in the other series. A particularly effective filtering-out of these variations would be necessary in order to see in the time domain (after a simple subtraction of series) the differences in amplitude response implied by the analyses in the frequency domain. Of course, the event-by-event comparisons previously mentioned are a subjective form of the kind of filtering needed and do give support to the response-type slope estimates.

For the large-scale slope records, fluctuations in wind stress account for 66–81% of the variability (Table 3). In general, the alongshore stress is more useful in explaining the variability. For the case of Cedar Key–St. Petersburg, cross-shore stress is more important, but here the use of Apalachicola winds and the chosen coordinate system is less applicable than elsewhere. Note that 81% of the variation in the Shell Pt.–Cedar Key slope (for which the chosen stress and coordinate system are most applicable) is accounted for by alongshore stress alone. In that case, a slope of about $15 \times 10^{-7}$ (sea level rising downwind) is produced by a 1 dyn cm$^{-2}$ alongshore stress. This value is comparable to the gain values determined from a cross-spectral analysis between Shell Pt.–Cedar Key slope and vector wind stress (Fig. 10). There, slopes of $(12\text{–}26) \times 10^{-7}$ per unit stress are found with the alongshore direction being clearly the most important.

## 6. Summary and discussion

The sea level response in the West Florida Bight has two important time scales: about 3.5 days and 5–10 days [the same as found by Marmorino (1982) for the entire West Florida Shelf response]. At 3.5-day periods, sea level responds to energetic cross-shore wind-stress forcing after a lag of some 6–10 h. At the longer periods, sea level is most coherent with the dominant alongshore stress, the time lag now being on the order of a day. On the average, the coastal sea level signal propagates at about 850 km day$^{-1}$ in the direction of movement of atmospheric cold fronts. Marmorino (1982) gives a speed of 3400 km day$^{-1}$ along the entire West Florida coastline but shows (his Fig. 7) that it is much less in the Bight. Cragg *et al.* (1982) have emphasized the clear response to alongshore stress and negligible response to cross-shelf stress; however, their data were filtered to suppress periods of less than 4 days, so the strong response to cross-shelf forcing in the Bight was apparently missed.

TABLE 3. Winter period regression analysis in the form: slope $\times 10^7 = c + a\tau(x) + b\tau(y)$. Units of $a$ and $b$ are (dyn cm$^{-2}$)$^{-1}$. Values in parentheses indicate insignificant (90% level) statistical relationships. $\Delta y$ is the distance between stations.

| Station pair | $\Delta y$ (km) | $c$ | $a$ | $b$ | Variability* $\tau(x)$ | $\tau(y)$ |
|---|---|---|---|---|---|---|
| A–CK | 17 | −2.18 | (7.4) | −67.3 | 0 | 34 |
| D–A | 52 | 2.31 | 16.8 | 13.2 | 14 | 21 |
| D–CK | 69 | 1.20 | 14.5 | −6.6 | 29 | 15 |
| CK–SP | 146 | 2.42 | 20.6 | 6.8 | 52 | 14 |
| Sh–CK | 158 | 0.75 | (0.8) | 15.2 | 0 | 81 |
| Sh–SP | 303 | 1.56 | 10.3 | 11.2 | 20 | 58 |

* This is the percent of the variability in slope that is accounted for by the wind-stress components. The smaller value gives the increase in accountable variability that results from including the secondary component. The total explained variability is the sum of the two values.

Previous studies (Cragg *et al.*, 1982; Hsueh *et al.*, 1982; Marmorino, 1982) have emphasized that the amplitude of the event-scale signal increases northward. This is, of course, consistent with dynamical models of shelf circulation which show that the low-frequency response to large-scale forcing depends only on the forcing acting over the region in the backward direction, backward meaning the direction from which free shelf waves propagate (Clarke, 1977; Csanady, 1978); hence, the response increases northward because there is an increasingly larger (backward) region contributing to that response. As a result, coastal sea level will rise downwind for an alongshore wind of either sense. This does appear to describe the large-scale response of the Bight during the winter events.

In Table 4, alongshore slopes from this present study (columns 3 and 4) are compared to previous results for station pairs of increasingly larger separations. (Again, only results from the longer winter study are used since these have greater statistical certainty.) Column 4 gives values of slope based on differences in the response shown in Fig. 6, but other measures of response (from spectral analysis, etc.) would do as well. Since the alongshore stress is the more energetic, these values can reasonably be associated with that stress orientation and compared to the other values obtained from data (columns 3 and 5). The agreement is not bad considering the different methods used to calculate slope (see below). The last column in Table 4 is based on results from a steady-state, $x$–$y$, barotropic circulation model for the West Florida Shelf forced with a unit alongshore stress (Marmorino, 1982). The model resolution in the $y$ direction is only 15 km so that finer details of coastline variation are excluded. However, the model does show (Marmorino's Fig. 14) a minimum in sea-level response just north of Cedar Key, apparently caused by the bend in the coastline there. As a result, the A–CK slope is negative (sea level decreasing in the direction of the wind), in agreement with the observations. This may be taken as some evidence for the creation of local changes in the sense of alongshore slope as a result of small-scale variations in nearshore bathymetry. [The model calculations of Hsueh *et al.* (1982) use a resolution of 28 km, but averaging in the model produces an effective value even larger; thus, their results are not expected to, and do not, show the small-scale slope reversal in the neighborhood of Cedar Key.] The one point of notable disagreement in Table 4 involves Station D and Cedar Key. There, the steady-state model and the response estimates of slope both give relatively small positive values, but the regression analysis that uses the direct calculation of slope (Table 3) gives a value of −7. However, since the amount of variability accounted for by the alongshore stress in that analysis was only 15%, the other values should probably be considered more realistic estimates of the alongshore-wind-driven slope.

TABLE 4. Summary of values of coastal sea-level slope (in units of $10^{-7}$) for an alongshore wind stress of 1 dyn cm$^{-2}$ in the positive $y$-direction.

| Station pair | $\Delta y$ (km) | Table 3 (Winter study period) | Fig. 6[1] (Winter study period) | Previous studies | Steady-state model[5] |
|---|---|---|---|---|---|
| A–CK | 17 | −67 | −11 | — | −5 |
| D–A | 52 | 13 | 7 | — | 3 |
| D–CK | 69 | −7 | 3 | — | 2 |
| CK–SP | 146 | 7 | 13 | 9[2] | 8 |
| Sh–CK | 158 | 15 | 3 | 7[3] | 7 |
| Sh–SP | 303 | 11 | 8 | 13[4] | 8 |

[1] Calculated as (differences between amplitudes in Fig. 6)/0.4$\Delta y$, where 0.4 dyn cm$^{-2}$ is the winter-period amplitude of the $y$-component fluctuating wind stress.

[2] From the 3-year study of Cragg *et al.* (1982), adjusted to unit stress and $\Delta y$ = 146 km.

[3] From Marmorino (1982), calculated as in (1).

[4] Actually Shell Pt.–Clearwater (open coast station nearest SP), but adjusted to $\Delta y$ = 303 km. From Marmorino's (1982) analysis of 1978 winter data.

[5] From Marmorino (1982), see his Fig. 14.

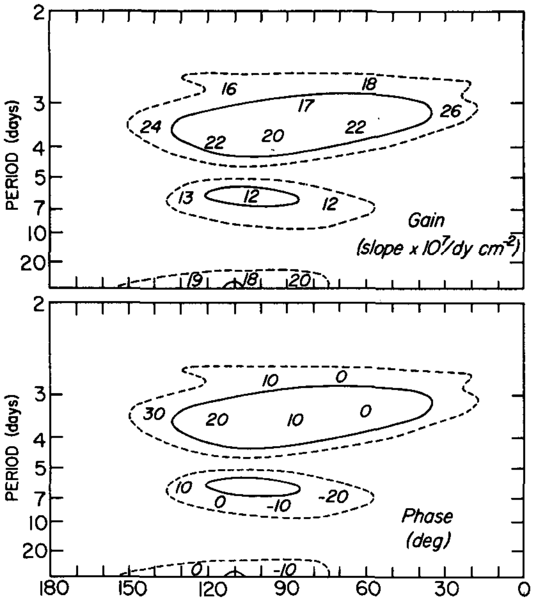

FIG. 10. Cross spectra between Apalachicola wind stress and sea-level slopes between Shell Point and Cedar Key during the winter study period. See Fig. 4 for explanation.

During the autumn period, a large response—about 60 cm (dyn cm$^{-2}$)$^{-1}$—was found in response to cross-shelf forcing. Corrected for atmospheric pressure variations and more realistic wind strength, this value may reduce by about half, say, to 30 cm (dyn cm$^{-2}$)$^{-1}$. The model by Hsueh *et al.* (1982) gives the response to an impulsively applied (and thereafter maintained) $x$-directed unit wind stress. After half a day or so, the modeled response is basically confined to the Bight in which sea level sets up much as in a lake, with a nodal line located nearly along the 30 m isobath (about 100 km offshore). Maximum response, about 20 cm, occurs near Cedar Key where the shelf is widest. Thus, the model implies a gain of about 20 cm $(\text{dyn cm}^{-2})^{-1}$, on the order of what is observed. Hayes (1979) has also found that cross-shelf winds can be quite significant in setting-up sea level over the inner shelf, though in his case, where the shelf was relatively narrow and deep, the amplitudes were smaller than found here. Recently, Chuang and Wiseman (1982) have studied in detail the direct cross-shelf wind set-up that is favored over Louisiana's very shallow inner-shelf. It would be interesting to apply a frictional model (such as they used) to the detailed nearshore bathymetry in the Bight to determine if an autumn-like response (with a relative minimum near Station D, where the isobaths lie closest to shore) could be reproduced under a cross-shelf wind.

Evidence has been presented for the existence of small-scale variations in the coastal sea-level slope in the West Florida Bight. During a period when the large-scale slope was clearly driven by events of alongshore stress, the small-scale slope—whether obtained directly (by differencing) or by isolating the wind-driven part of the signal in some way—was found to agree or disagree with the sense of the large-scale slope. It is thought that slope reversals arise through the influence of variations in the very nearshore bathymetry and changes in coastline orientation, and, in the case of direct calculations, by contamination of the desired signal by lower-frequency non-wind-driven fluctuations. The dynamical importance of the variations may lie in their extension offshore to produce cross-shelf mass exchange as part of a kind of "shelf circulation cell" (Csanady, 1981; Pettigrew, 1981). In a comprehensive investigation of the dynamics of the inner 12 km (water depths $\leqslant 33$ m) of the wide shelf off Long Island, Pettigrew (1981) has observed episodes of significant cross-shelf transport for which a balance of the term $fu$ in the depth-averaged alongshore equation appears to require a local alongshore sea-level slope (of roughly 1 cm in 30 km) that is very much different from the large-scale slope estimated from stations 200 km apart. The examination of slope over small scales may thus serve to reveal locations where cross-shelf flows are likely to occur.

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