Wind-Induced Sea-Surface Slopes on the West Florida Shelf

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ABSTRACT

Tidal and meteorological records at stations in the eastern Gulf of Mexico have been studied. The sea-level response is a maximum for winds along the coast and varies symmetrically with angle. The coherence is maximum at periods of 4–10 days. The horizontal coherence of sea level is high out to 500 km for 4–10 day periods. The horizontal coherence for wind (measured at coastal stations) is high out to at least 500 km. The amplitude of the response of sea level to winds is larger by a factor of 4 here, on a broad shelf, than on the Oregon coast, which is more narrow by about the same ratio. A response of \( \sim 16 \) cm is induced by \( \sim 4 \) m s\(^{-1}\) wind. This response, or transfer function, is uniform over the spectral range (4–100 days). The sea-level response to the longshore wind stress is not linear, but to the power 0.8 \( \pm \) 0.1, and is attributed to the relatively low tidal currents in this region. The large horizontal coherences of wind and sea level imply broad longshore flows extending 500 km or more along the coast. Over 85% of the variance between 4 days and 3 years is contained in fluctuations with periods less than 3 months. A longshore slope of sea level is observed; in the 4–10 day band this slope can be explained by longshore variation in the width of the shelf. A mean longshore slope of \( \sim 0.6 \times 10^{-3} \) is found, and it may be caused by the (weak) mean winds. Freely propagating coastal trapped waves are found in a narrow band near 0.18 cpd.

1. Introduction

A detailed analysis of coastal circulation is difficult because so many parameters influence and describe the flow. As part of a long-range study of coastal currents on the west Florida continental shelf, we have examined one of the important driving forces—the wind—to determine its correlation with sea level at the coast, with the idea that sea-level variations are a good measure of longshore current fluctuations. Many studies have shown that sea level is highly coherent with longshore currents at the periods associated with synoptic-scale wind events, from \( \sim 2 \) to 20 days (Kundu et al., 1975; Smith, 1974; and others). Models containing wind stress and bottom stress are able to explain many of the observed features (Sandstrom, 1980; Noble and Butman, 1979). The development of ideas on circulation on continental shelves has been reviewed by Beardsley and Boicote (1981), and a series of papers by Csanady (1981) provides a timely overview. The purpose of this paper is to summarize our initial results concerning the wind-induced response of coastal sea level along the West Florida Shelf. It is a wide shelf, with small tidal currents (compared with the U.S. east coast) and has for the most part relatively weak wind forcing.

Ichihye et al. (1973) computed monthly charts of surface currents and Ekman transports in the eastern Gulf of Mexico. Their analysis shows the surface currents to be highly variable in both direction and speed from month to month; no clear seasonal coastal patterns emerge. The atlas of wind-stress values by Hastenrath and Lamb (1977) also includes the Gulf of Mexico. Drift-bottle studies off the northwest Florida coast by Tolbert and Salsman (1964) suggest that the primary mechanism of surface-water transport in this area is the wind. The first work of our knowledge that included observations of currents in this region was by Niiiri (1976). Further data on current measurements have been presented by Kobliinsky and Niiiri (1980) by Marmorino (1982, 1983), and by Mitchum and Sturges (1982).

The coherent width of the wind field is frequently larger than the offshore extent of the current, even on a broad shelf, and so this scale is usually ignored. It is assumed that the width of the continental shelf is an important offshore scale, as well as the frictional scale. We devote a major part of this manuscript to determining the horizontal scales over which the wind forcing and the coastal response are coherent, as we have been unable to find such results for this region in the literature. This paper is organized by sections as follows: 2. Data; 3. Horizontal wind coherence; 4. Horizontal sea-level coherence; 5. Sea-level response to wind; a. Sea-level response, b. Variation of wind direction; 6. Wind-induced longshore currents; 7. Coastal trapped waves; 8. Longshore pressure gradient.
2. Data

Tide gauges used for this study are shown in Fig. 1. The continental shelf is broad; the 100 m isobath can be at least 200 km offshore. The gauges at St. Petersburg and Pensacola are located inside bays. The Cedar Key gauge is exposed on an open coast. The remarkable similarity we find between low-frequency sea-level records at St. Petersburg and Cedar Key, however, suggests that small-scale topography surrounding each gauge is relatively insignificant, as has been found earlier (e.g., Groves, 1957).

Three years of hourly heights for 1965 through 1967 were analyzed at each tidal station. Daily atmospheric pressure and resultant winds, computed from eight observations per day, were obtained from the Tampa, Tallahassee and Pensacola airports. The three-year interval was chosen for the absence of gaps in all three records. The diurnal and semidiurnal tidal components were removed by low-pass filtering, to yield daily values. The filter is designed to pass energy having periods greater than 4 days; periods less than 2.8 days were suppressed to less than 0.1%. Wind values were similarly filtered.

3. Horizontal wind coherence

To determine whether or not winds recorded at the various airports adequately represent the wind field over the distances between them, it is necessary to know the horizontal coherence of winds in this area. Frankignoul and Müller (1979) provide a summary of information about the scales of wind forcing over the open ocean. The results shown here are necessary to provide more detail about the local meteorology. Figs. 2 and 3 show the spectra of winds at Tampa and Tallahassee, which are 320 km apart. In the spectral range of 4–10 days the power density rises approximately as $f^{-1.5}$. This rise in power, associated with the passage of frontal systems, has been noted in other wind spectra (Van der Hoven, 1957; Oort and Taylor, 1969). Of the total kinetic energy contained in periods of 4–100 days, approximately 50% is found in this initial "cyclone rise". The spectral range 10–100 days is characterized by a relatively uniform level in the north–south component, and by a gradual rise in power in the east–west.

A variety of correlations for wind components at several separations is shown in Fig. 4. We have treated the separation between Tampa and Tallahassee as though it were strictly north–south (a ~35° error) for ease of using standard north–south and east–west wind components in determining transverse and longitudinal coherence functions (only for the purpose of this figure and not in the subsequent analysis). The shortest distance (in Fig. 4) is the Tallahassee–Pensacola separation while the slightly longer one is the Tampa–Tallahassee separation. As is clear from Figs. 2 and 3, the coherence is relatively uniform over the band from 4 to 100 days. Fig. 4 shows that winds are highly coherent over distances greater than the width of the shelf, and that the transverse and longitudinal correlations do not differ significantly over the first 250 km. The availability in recent years of data from an offshore weather buoy at 26°N, 86°W has shown that although the winds are stronger offshore (as was well known), the low-frequency variability is very similar to that seen at Tampa (e.g., Marmorino, 1983; New England Coastal Engineers, 1981). The coherence values here fit within

![Fig. 1. Map of the northeastern Gulf of Mexico showing tide gauges (triangles) and weather stations (circles) used in this study.](image-url)
the scatter of points in Noble and Butman's (1979) Fig. 5.

The weather systems on the West Florida coast generally move from northwest to southeast. At periods of 4–10 days Pensacola north–south wind components lead Tallahassee and Tampa north–south wind components by 5 and 21 h, respectively, on the average. East–west components remain surprisingly coherent and in phase over the whole northeastern Gulf of Mexico.

Longshore winds at Pensacola, which are approximately east–west, lead longshore winds at Tampa, which are approximately north–south, by 15–20 h in periods of 4–10 days. This "longshore coherence" averages 0.35 and is as high as 0.6 at 0.18 cpd (an energetic region in the wind spectrum) as shown in Fig. 4. Although phenomena at higher frequencies may have smaller scales (e.g., the land-breeze, sea-breeze effect), we are concerned here with events associated with the passage of large weather systems.

From these calculations we conclude that winds recorded at Tampa, Tallahassee, and Pensacola represent well the wind field over the northwest Florida Shelf in the frequency band 0.25–0.01 cpd at distances out

![Graph showing power spectrum and cross spectrum of Tampa-Tallahassee north-south winds.](image-url)
to about 200 km or more. With these results, therefore, Cedar Key sea levels can be justifiably paired with Tampa winds in subsequent analyses.

Daily atmospheric pressure values at Tallahassee and Tampa were also compared. The coherence was uniformly above 0.8 over periods of 4–100 days; daily pressures differed by only 1.4 ± 0.9 mb between the two stations. Therefore, daily pressure for Cedar Key was linearly interpolated from Tallahassee and Tampa values for the analyses that follow.

4. Horizontal sea-level coherence

The filtered sea levels were adjusted for the inverted barometer response to 1014 mb (the mean atmospheric pressure for the three stations, 1965–67). This removal of pressure-induced sea-level fluctuations reduced the total variance of the sea-level record, periods 4 days to 3 years, by 25%.

The barometer factor as a function of frequency is computed as the ratio of sea-level and atmospheric pressure. This “in phase” barometer factor, computed from Tampa atmospheric pressure and St. Petersburg unadjusted sea level, is shown in Fig. 5. At periods of 4–20 days, the barometer factor has an average value of 2.1 ± 0.3, obviously being enhanced by the effect of coastal currents. In the vicinity of 20–30-day periods, the barometer factor approaches the isostatic value, a result also found by Mysak and Hamon (1969) using records on the North Carolina coast. From other results presented later in this paper, we know that longshore wind and sea level lose coherence in the frequency range near 0.05 cpd. Chelton and Davis (1982) have examined the sea-level response to atmospheric pressure fluctuations. They concluded that a major part of the over-response along the California coast (at low frequencies) is from forcing farther down the coast, to
shown in Fig. 7. Sea-level fluctuations—which are presumably related to longshore currents—show coherence out to at least 500 km. The highest sea-level coherences occur over the most energetic part of both wind and sea-level spectra at periods of 4–10 days. As the coherence between longshore winds and sea level falls—in the vicinity of 10–20 days—sea-level fluctuations lose correlation in a similar manner. At low frequencies (seasonal, rather than synoptic), the scales seem to get much longer. Enfield and Allen (1980) found that sub-annual signals are coherent for many thousands of kilometers along the U.S. Pacific coast. Chelton and Davis (1982) found coherence in sea-level variations (their Fig. 4a) at periods both longer and shorter than a year from Mexico to Alaska.

5. Sea-level response to wind

Spectra of longshore winds and sea level at Tampa are shown in Fig. 8. The adjusted sea-level spectrum and the cross spectrum between sea level and longshore wind are as expected very similar. The rise in sea-level power at periods 4–10 days has the same slope, $f^{-1.5}$, as found in the wind spectra. Of the total variance contained in sea-level signals of periods 4–100 days, approximately 50% is found in this initial 4–10 day rise.

Except for frequencies below 0.05 cpd, the coherence between adjusted sea level and longshore winds at Tampa is uniformly high, and the phase shift is less than 15°. Cross-spectral analysis of Cedar Key and Pensacola sea levels and longshore winds yielded similar results.

The “in phase” transfer function from this calculation (the ratio of the co-spectrum of sea level and longshore winds to the auto-spectrum of longshore winds

![Graph showing barometer factor as a function of frequency](image_url)

**Fig. 5.** In phase barometer factor (cm mb⁻¹) as a function of frequency; computed from unadjusted sea levels at St. Petersburg and atmospheric pressures from Tampa. Dashed line represents the isostatic value of −1.0 cm mb⁻¹.
winds) is fairly uniform over the spectral range of 4–100 days, at 3.9 ± 0.3 cm (m/s)^{-1} (in all bands where there is power reasonably above the noise). This uniformity suggests that there are no preferential frequency bands for the efficient transfer of momentum from wind to water, consistent with present ideas that the response of coastal currents to wind is highly damped by bottom friction. The coherence decreases uniformly for periods longer than 10 days, and (for this three-year data segment) the two signals are not significantly correlated (at 99% level) at 100 days; the phase, however, also shows a consistent change. The 4–10 day events force an almost instantaneous, barotropic response. For periods longer than 3 weeks (f < 0.06 cpd), the phase lag grows steadily, and suggests that a larger phase delay, perhaps associated with shelf waves, may hold for longer periods.

By averaging the phase differences at frequencies where both coherence and power are high, we have examined the delay between adjusted sea level and longshore winds. Sea level consistently lags winds by 10 h, over periods of 4–10 days, at St. Petersburg. Cedar Key shows a similar lag (8 h, but since the winds are from Tampa, the difference between the 8 h and 10 h results are probably merely phase shifts in the wind). A surprising result, however, is that the delay at Pensacola is 24 h, or twice as long. The delay between sea levels at Pensacola and at Cedar Key or St. Petersburg, however, is only 4 h (at periods of 4–10 days).

The explanation for this effect probably lies in the fact that wind systems travel from northwest to southeast along this coast, so the wind begins to act first at (of these stations) Pensacola (see, for example, Marmorino, 1982). Daily sea levels at the three tide gauges have been adjusted as appropriate for these lags in subsequent calculations by changing the time origin of the sea-level records.

A plot of lag-corrected, adjusted sea level for St. Petersburg and daily longshore wind speed at Tampa is shown in Fig. 9. These January data show the typical high correlation that exists between low-frequency sea-level fluctuations and longshore winds.

a. Power law of sea-level response to wind stress

In the original course of this work a log–log fit of sea-level response vs the longshore component of wind indicated that sea level was not responding linearly to wind stress, but to a power significantly less than unity (Cragg and Sturges, 1974, Fig. 9). Subsequently, a careful analysis of more recent data at St. Petersburg (wind) and Tampa (sea level) was carried out to explore this interesting finding.

Prior to the determination of the power law exponent, an extensive set of numerical experiments was done using artificially generated data in order to find the most reliable method of fitting the assumed power-law relationship to the data. This was found to be necessary because the effects of random variability in the data can produce a bias in the power-law analysis. That is, the log–log fit requires that sea-level and wind be of the same sign whereas in nature they do not always satisfy this requirement. Neglecting such points or averaging the data sets in some manner is thus required and introduces the possibility of bias.

A number of important results emerges from the numerical experiments (discussed by Mitchum and Sturges, 1983). The two most important findings are:

1) A “batch averaging” of the data, in a manner similar to that used by Scott and Csanady (1976) in determining a bottom-friction parameter off Long Island, decreased the scatter greatly.

2) Separating the data into year-long subsets, determining the exponent for each year, and then averaging these as independent determinations also permit significant improvement.

We are able to obtain an accurate fit to the artificial data with peak signal-to-noise ratio (SNR) lower than 5. Although the noise in the real series may not be
**Figure 7.** Horizontal coherence-squared of adjusted sea level along the West Florida coast. Coherences are averaged over periods of 4–10 days and 10–50 days at each pair of tide gauges. The 99% confidence limit is shown by the dashed line.

**Figure 8.** Power spectrum of St. Petersburg adjusted sea level and cross spectrum with longshore winds from Tampa, computed from 1000 daily values. The 95% confidence limit on coherence-squared is shown by the dashed line.

Uniform, or random, as our experiments assumed, we do not expect a SNR this low.

The sea-level data are corrected for the inverted barometer effect. Wind stress is calculated using a drag coefficient that varies linearly with wind speed (Wu, 1980). Assuming that sea-level variations at the coast, \( \eta \), are proportional to wind stress to some power \( \alpha \), we write

\[
\eta = \tau^\alpha.
\]

From five years of data, the power-law exponent \( \alpha \) is found to be \( 0.8 \pm 0.1 \). A nonlinear least-squares technique (Hartley, 1961) using these data yields a result identical to the earlier log-log fit discussed above. To test for seasonal variability, including stratification in the water column, the data are separated into overlapping two-month groupings and the exponent determined for each subset (Fig. 10). The standard deviations are larger because there are fewer data points. Although the winter–summer difference is only at the level of one standard deviation, the result is suggestive.

In order to understand a nonlinear sea-level response to wind forcing, it is natural to look for a term in the momentum equations that is nonlinear and sufficiently large to cause observable effects. The nonlinear terms in \( \mathbf{u} \cdot \nabla \mathbf{u} \) are seen, by simple scale analysis, to be small and will be neglected. The alongshore component of the instantaneous bottom stress term, which is the leading candidate remaining, can be written

\[
\tau_b^V = C_D |\mathbf{u}| v,
\]

where \( |\mathbf{u}| \) is the current magnitude, \( v \) the alongshore
velocity component and $C_D$ a constant drag coefficient. This (nonlinear) bottom stress term is known to be significant in the momentum balance on the West Florida Shelf (Mitchum and Sturges, 1982). In the nearshore region, the approximate balance in the synoptic band is between low frequency wind and bottom stresses, consistent with the theory of Csanady (1978). Thus, we write

$$\tau \approx C_D |u|v,$$

(3)

where $\tau$ is the low-frequency alongshore wind stress and $|u|v$ is the low-frequency part of $|u|v$.

It is instructive to examine two limits of (3) to demonstrate how the bottom stress term in (3) may explain the observed power law. The first limit is the situation in which the high frequency currents (here, tides) are much more energetic than the low-frequency ones. Csanady (1976) showed that in this case

$$\tau' \approx rv,$$

(4)

where $r$ is a constant friction parameter equal to the drag coefficient times a high frequency rms current. Observations (e.g., Mitchum and Sturges, 1982) and numerical models (e.g., Hsueh et al., 1982) show to good approximation

$$v = \gamma \bar{\eta},$$

(5)

where $v$ is the low-frequency, alongshore current, $\eta$ the low-frequency coastal sea-level fluctuation and $\gamma$ a constant of order 1. Eqs. (4) and (5) thus imply that sea level and wind stress should be linearly related.

In the other limiting case, the high-frequency currents are much less energetic than the low-frequency currents. In this case, (2) shows that bottom stress varies approximately as alongshore current squared. Again, use of (5) implies that sea level should vary as the square root of wind stress in this limit.

These limits show that, depending upon the relative magnitudes of the high- and low-frequency currents, the power-law coefficient should be between 0.5 and
1. The observed value of 0.8 ± 0.1 is nicely consistent with the fact that on the West Florida Shelf the high-frequency (tidal) and low-frequency (synoptic) currents are comparable in magnitude, with the high-frequency components being slightly larger (Mitchum and Sturges, 1982).

The finding that sea level does not respond linearly to alongshore wind stress can have important consequences for some types of data analysis. An example is an attempt to adjust a sea-level series for synoptic wind forcing in order to examine variability which is not wind-forced.

b. Variation of wind direction

To avoid contaminating the shorter period wind and sea-level correlations it seemed appropriate, for some parts of this work, to remove the annual cycle from the sea-level record. The sea-level series were Fourier decomposed, and the 12-, 6- and 4-month terms were suppressed; this operation eliminated from 30% to 45% of the total variance of the adjusted sea-level records. The wind data were similarly filtered.

One way to examine wind-induced sea-level response is by a sorting procedure, to correlate winds of varying speeds and directions with the corresponding adjusted sea levels at each gauge. Adjusted sea level at Cedar Key versus wind direction is shown in Fig. 11 [and for the other stations, by Cragg and Sturges (1974)]. The maximum rise and fall of sea level occurs when winds are approximately parallel to the average coastline direction. Onshore–offshore winds do not produce sea-level fluctuations that are reliably above the noise level. The standard deviation of each data point is on the order of 2 cm. The rise and fall of sea level at St. Petersburg, as a function of wind direction, has roughly the shape of a sine wave. At higher wind speeds, the effects of wave set-up will get worse (e.g., Thompson and Hamon, 1980). The rise and fall of sea level (Fig. 11) is approximately symmetrical with respect to wind direction, but there is an asymmetry of ~15% in peak amplitudes. On the New England shelf, Beardsley and Butman (1974) found the response to be asymmetric. Noble and Butman (1979) find the response to be symmetric in the Middle Atlantic section of the U.S. east coast, and describe mechanisms for the difference. In addition to the relative maximum associated with longshore wind, there also seem to be influences of the coastline orientation on smaller scales. The maxima in Fig. 11 at ~200° and near 15° to 30° are, we assume, associated with the coastline indentation just north of Tampa.

We have treated each day’s wind and sea-level data as independent, in determining error bars. This error analysis is precise for 4-day periods, but for the lower-frequency motions this procedure underestimates the error. For 20-day periods the correlation function goes through zero at 5 days; the error bars shown would be too small by ~2. As half the variance is found between 4 and 10 days, however, the reader may mentally treat the error bars as seems fitting.

The similarity in shape of all the curves for all three locations indicates that the wind-induced sea-level response is not primarily a local effect occurring in the immediate vicinity of a gage. The horizontal coherence of the wind field suggests that a broad longshore geostrophic flow may extend as far as from St. Petersburg around the coast to Pensacola, a distance of over 500 km.

Our average “in phase” transfer function between sea level (cm) and longshore winds (m s−1) has a value of 3.9 ± 0.3. Because the transfer function between sea level and winds is fairly uniform over frequency, and the magnitude of the quadrature spectrum is small, the transfer function is comparable with the linear regression coefficient between the two series. For a given wind speed (<5 m s−1), the sea-level response on the west Florida shelf is about four times as large as the fluctuations recorded off Oregon (Smith, 1974). This ratio is presumably related to the shelf width, and is discussed briefly in a later section.

6. Wind-induced longshore currents

It seems appropriate to point out that the observed sea-level departures are consistent with known currents. From Fig. 10 we see that a 4.5 m s−1 wind will induce a sea-surface fluctuation Δh of ~20 cm. If we assume that a representative geostrophic surface current speed is 15 cm s−1 (i.e., a typical wind factor of 3%—a result also obtained by Sandstrom, 1980), the width of a
uniform geostrophic current is on the order of 170 km. These values are all internally consistent.

7. Coastal trapped waves

Continental shelf waves have been discussed for many coasts; summaries are given by Clarke (1977) and by Mysak (1980). We have computed spectra of adjusted and unadjusted sea level, longshore winds and atmospheric pressure. In our examination of the spectra, all the peaks in elevation corresponded to similar ones in the longshore wind.

However, the phase relationship between St. Petersburg and Cedar Key shows that Cedar Key consistently leads St. Petersburg by 2–4 h, over a frequency band of 0.4 to 0.5 cpd. This phase shift is the wrong sign for coastal trapped waves. Similar cross-spectral analyses of the other sea-level records show small phase shifts, over most frequencies, that have the same (wrong) sign. Because the weather systems in this area travel from west to east, however, the forced sea-level response dominates the sea-level records. The present ideas of wind-forced response on the shelf suggest that the primary momentum balance is between wind stress and bottom stress. Our results (based on a very unsophisticated search) suggest that any free waves present have amplitudes substantially smaller than the forced waves. This work is being continued, and will be reported separately.

There is a lag of 10 h, at 0.18 cpd, in the Pensacola–Cedar Key cross spectrum, however, which is the correct sign for coastal trapped waves, at a frequency where the wind spectrum also has energy over a broad scale. This delay suggests a phase speed of $\sim 300 \text{ km day}^{-1}$.

This is too fast, by a factor of about 4, for a free barotropic continental shelf wave on this coast but no correction has been made for the presence of forced waves. Enfield and Allen (1980), however, find coastal trapped waves propagating poleward along the U.S. Pacific coast (at much lower frequencies) at speeds closer to these.

8. Longshore pressure gradient

The sea-surface slope along the coast was determined by sorting the sea-level differences between tide gauges versus wind speed and direction. The difference between St. Petersburg and Cedar Key as a function of wind direction is shown in Fig. 12. The zero-level on the ordinate, as before, is the mean difference between the two gauges in times of very light winds; i.e., all wind components $< 1 \text{ m s}^{-1}$.

The response curves at each station (Fig. 11) are only approximately symmetrical, and when the longshore slope is computed, the asymmetries become more pronounced. Fig. 12 shows that the longshore slope is much larger for wind from the southeast than for winds from the northwest. It turns out, however, that this slope is not merely the result of an asymmetry at Cedar Key (Fig. 11) subtracted from a more nearly symmetrical response at St. Petersburg. The response at Cedar Key is larger in almost every wind direction than at St. Petersburg. The longshore slope is shown (Fig. 12) for the cases where winds were $4.5 \text{ m s}^{-1}$, as having the largest amplitudes and perhaps the best signal-to-noise ratio. For cases of winds $\sim 3.5 \text{ m s}^{-1}$, however, the alongshore slope shows a maximum that is smoother (with respect to direction) and is $\sim 5 \text{ cm}$.

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**Fig. 12.** Longshore slope of sea level on the West Florida shelf versus wind direction. Sea-level-difference cases between St. Petersburg and Cedar Key gauges are computed for winds that average $4.5 \text{ m s}^{-1}$. Standard error bars average 1 cm.
in the range 135–160°. The maximum slope occurs when the wind is approximately parallel to the 40 m isobath (see Fig. 1).

For a longshore wind speed of 4.5 m s\(^{-1}\) the longshore slope has a magnitude O(3.0 \times 10^{-7}). The direction of the slope is such that sea level rises downwind. This slope is open to explanation by several independent mechanisms. The most likely explanation is that the width of the continental shelf changes (slowly) in the longshore direction. We can define a gain factor \(G\) as the ratio of coastal sea-level response to the wind. If we assume that the average longshore current can be represented as being approximately linear (for simplicity) in longshore wind speed \(W\) with a dimensionless proportionality constant, the “wind factor” \(\alpha\), then we may write, assuming geostrophic flow,

\[
G = f\alpha X g^{-1} = AX.
\]

Note, however, that the assumptions relating the wind to sea level and current drop out of (6) (except for any difference in the power laws; we assume the coefficient \(A\) could be a weak function of wind speed).

Using values from this paper for St. Petersburg, we estimate (using linear wind laws) \(A \approx 4 \times 10^{-4} \text{ s km}^{-1}\). Between St. Petersburg and Cedar Key, the width of the shelf increases 20%, from \(\sim 150\) to \(\sim 180\) km. For a wind speed of 4.5 m s\(^{-1}\), the result for \(A\) gives a calculated slope of \(\sim 4 \times 10^{-7}\) which is entirely consistent with the observed slope. (Note that this simple result uses the sea level response at St. Petersburg but no information about the longshore slope.) Thus, we find the result that the longshore change in shelf width may give rise to a longshore pressure gradient under the influence of a longshore wind. [A slightly different approach would be to use the values in Noble and Butman’s (1979) Table 3 to find an alternative value for the coefficient \(A\).]

Because separation between these two tidal stations (150 km) is much less than the coherence scale of the wind field, we have assumed that the wind field is approximately uniform between them. In many other situations involving long coastlines, slopes are found that are coherent with wind, but the wind stress changes substantially over these longer distances (Chelton and Davis, 1982; Enfield and Allen, 1980). Beardsley and Winant (1979) discuss forcing by freshwater runoff and open-ocean forcing. We assume that both are small here, and not coherent with local winds. The reasons for the asymmetry of the observed response presumably is related to the detailed shape of the coastline—but we offer no detailed explanation here.

To determine whether the sea-level fluctuations are coherent with wind at lower frequencies, Fig. 13 shows the seasonal data as a function of longshore wind component (here, taken north–south). There appears to be a reasonable correlation. Fig. 13 suggests that a weak mean wind may thus give rise to a longshore slope of \(\sim 0.6 \times 10^{-7}\) between the two stations. The slope of the line, however, is about \(-3.5 \text{ cm (m s}^{-1})^{-1}\), or three times as large as the slope inferred from the higher-frequency data in Fig. 12. As the mean value at longer periods is partly the residual of many shorter period events of higher intensity, the apparently higher gain factor may be largely the result of the averaging frequency.

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