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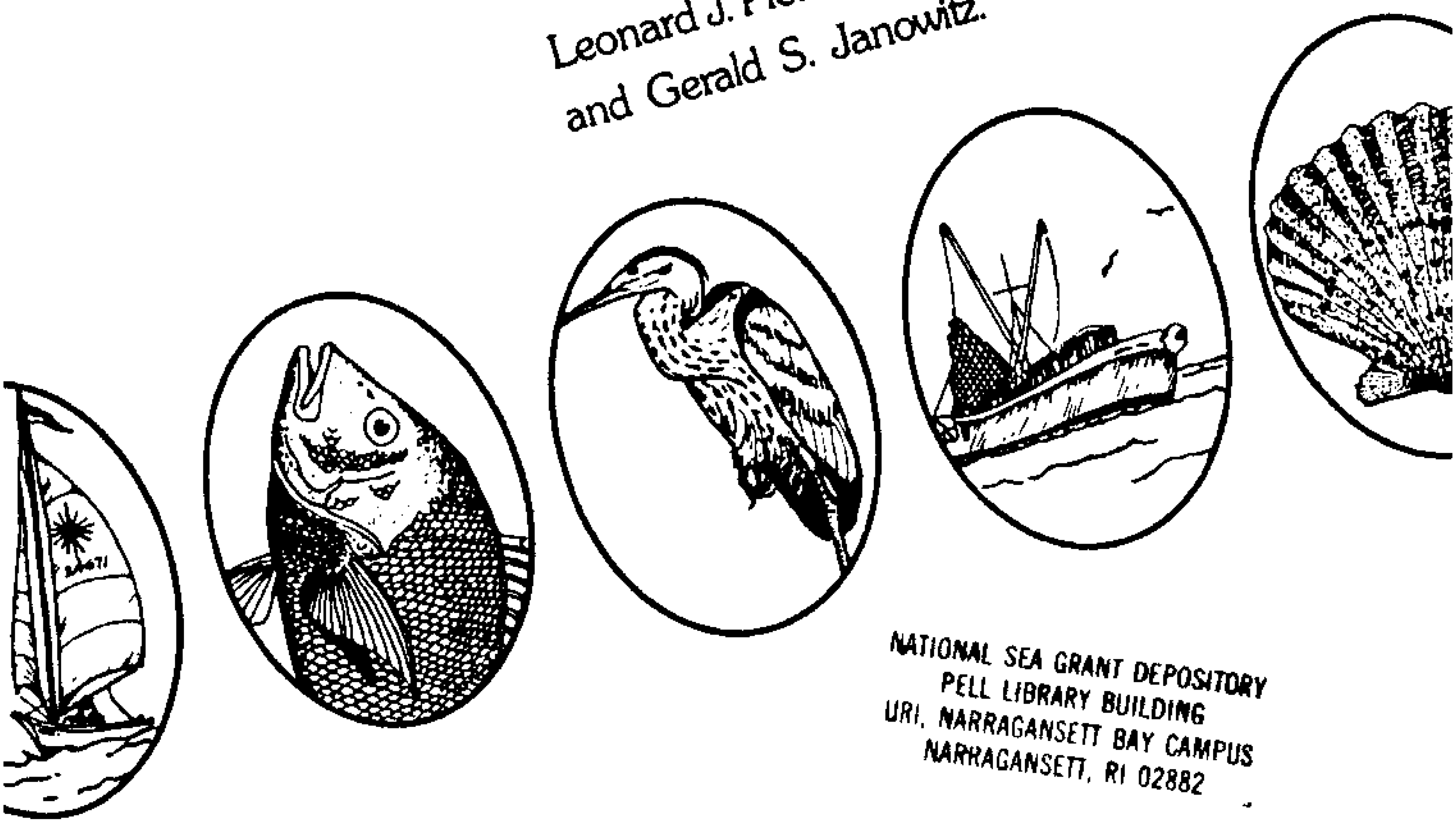

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THE VARIABILITY OF SEA LEVEL IN THE CAROLINA CAPES

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Introduction

Within the overall study of the dynamics Carolina Capes region of the South Atlantic Bight, (SAB) including Raleigh, Onslow and Long Bays (cf. Figures 1 and 2), is the study of coastal sea level and its relationship to atmospheric forcing. This report is a study of two four-month blocks of data collected during combined studies of the dynamics of the Carolina Capes shelf and the Cape Fear River Estuary. These study periods, 1 Sept.-31 Dec. 1974 and 6 June-27 July 1975, have been found to be representative of the subtidal frequency fluctuation "weather" of coastal sea level. The word weather is used to categorize several-day to several-week period fluctuations in sea level which ultimately coalesce to comprise the seasonal to yearly climatology of the regional coastal sea surface variability. For a more general description, refer to Pietrafesa et. al. (1978a)

The Response of Sea Level to Atmospheric Forcing: A General Commentary

The response of coastal surface elevation to continental atmospheric forcing has been examined by a number of investigators. Miller (1957 and 1958) studied the New England and New Jersey coasts; Hamon (1962, 1963 and 1966) studied the Australian coasts; Panshin (1967), Mooers and Smith (1968), Cutchin and Smith (1973), Smith (1974), Kundu, Allen and Smith (1975) and Huyer, Hickey, Smith, Smith and Pillsbury (1975) investigated the Pacific Northwest coast; Mysak and Hammon (1969) partially studied the North Carolina coast; Cragg and Sturges (1974) did an in-depth study of the West Florida shelf; while Brooks and Mooers (1977) studied the East Florida shelf.

Sea surface elevations along coastlines are related to both alongshore and cross-shore winds. Simple Ekman theory (Ekman, 1905) and subsequent studies of both set-up and set-down (Hidaka, 1953; Welander, 1957) and shelf wave generation (Hamon, 1962, 1963 and 1966; Mysak and Hamon, 1969; Cutchin and Smith, 1973; Huyer et. al., 1975) support the evidence for sub-inertial frequency correlations between atmospheric forcing and sea level signature.

While evidence for the existence of continental shelf waves has been inferred from the cross-statistical analyses of tidal and meteorological data, there have been few current meter observations made at a sufficient number of alongshore stations to confirm the existence of these phenomena. The recent work performed using moored current meter arrays between Cape Hatteras, N.C., and Charleston, S.C., at the direction L.J. Pietrafesa, will eventually sort out the various stable and unstable wave modes present in the Carolina Capes region. It is clear from initial reports of the current meter data results (Pietrafesa, 1977 and

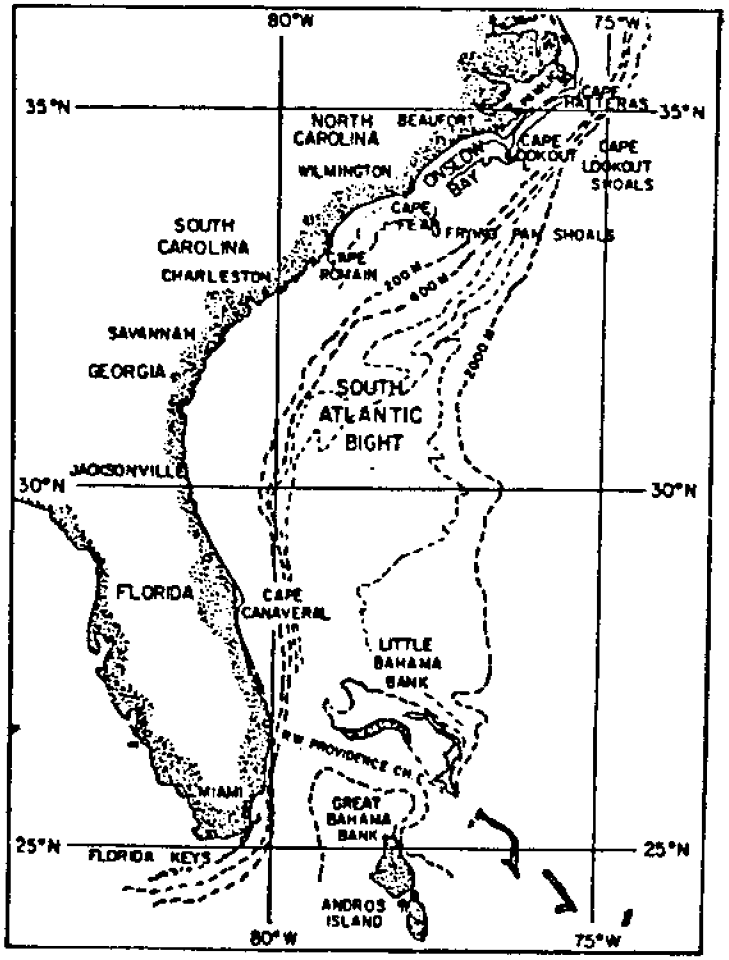


Figure 1: Map of the South Atlantic Bight

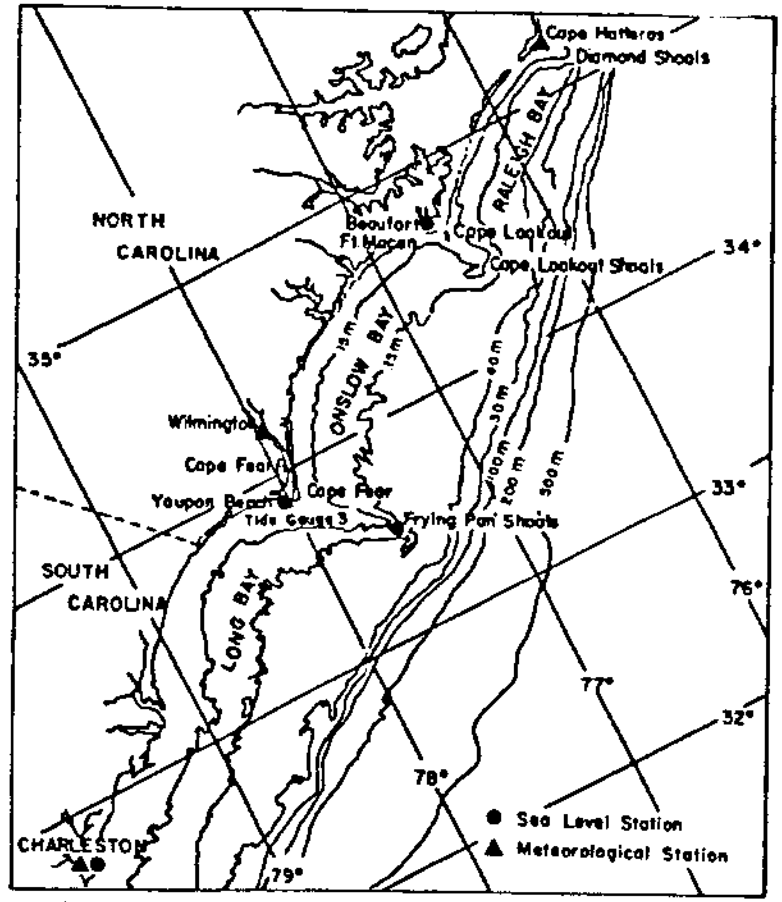


Figure 2: Map of the Carolina Capes

1978) that no southerly propagating continental shelf waves are evident between Frying Pan Shoal and Cape Hatteras. It is of note that Huyer et. al. (1975) confirmed the existence of such phenomena along the Oregon-Washington coast.

Wunsch et. al. (1969) determined that astronomical tides were responsible for most of the sea level variations along the eastern seaboard of the U.S. Nonetheless, a substantial fraction of the total energy exists in the low frequency (0.5-0.08cpd) fluctuation range. Events with fluctuation frequencies within this range are observed features of both the marine atmosphere and of the Gulf Stream front. With regard to the Gulf Stream, Pillsbury (1891), Parr (1933), Schmitz and Richardson (1968), Duing (1975), Duing, Mooers and Lee (1977), and Lee and Mayer (1977) observed such lateral, onshore-offshore periodicities of the current regime in the Gulf Stream off of the Florida Coast. Off Onslow Bay, Webster (1961) found 0.14 cpd cross-shelf Gulf Stream frontal meanders, which seemed to be correlated with the cross-shelf wind component. Orlanski (1969), Niiler and Mysak (1972) and Orlanski and Cox (1973) suggested that inherent baroclinic instabilities of the Gulf Stream could be responsible for the energy that appeared in the 0.5-.08 cpd range. These instabilities would subsequently force a shelf water response, which could appear in the signature of sea level. All of these possibilities will be examined in the analysis of sea level to follow.

Meteorological stations chosen for this study are: Cape Hatteras (HAT), Wilmington (WIL), Charleston (CHS), and Savannah (SAV), Sea level (i.e. tide gauge) stations include Beaufort (BFT), the Brunswick Tide Gauge No. 3 Intake gauge (BRI), 11 miles upstream from the Cape Fear River mouth, at the Cape Fear River mouth (CFM), Beaufort Inlet (BI), Frying Pan Shoals (FPS), Wilmington (WIL) and Charleston (CHS). These are located in Tables 1 and 2 and in Figure 2. Throughout this area the continental shelf tends to be shallow and generally broad, with the major exception of a narrowing at Cape Hatteras and excursions of shoals seaward of Cape Hatteras, Lookout and Fear (cf. Figure 2). The shelfbreak is typically at the 75 meter isobath. Though the topography in the vicinity of each gauge varies considerably, the similarity between the low frequency, low passed sea level records from the various stations supports the assumption that there is minimal location influence apparent in the spectral ranges considered, save for the Frying Pan Shoals and Wilmington data, which may be geographically and/or topographically influenced.

In this report, two four-month blocks of hourly heights, recorded to the nearest 2-3 centimeters, are analyzed at the sea level stations. Three-hourly observations of atmospheric pressure and wind speed and direction were obtained from the

Table 1

Meteorological Stations

Station	Latitude	Longitude
Cape Hatteras, N.C. (HAT)	35°16'N	75°33'W
Wilmington, N.C. (WAL)	34°14'N	77°57'W
Charleston, S.C. (CHS)	32°14'N	79°56'W
Savannah, Ga. (SAV)	32°05'N	81°06'W

Table 2

Tide Gauge, Sea Level Stations

Station	Latitude	Longitude
Beaufort, N.C. (BFT)	34°43.2'N	76°40.2'W
Wilmington, N.C. (WIL)	34°13.6'N	77°57.2'W
Frying Pan Shoals, N.C. (FPS)	33°29.1'5	77°35.4'W
Charleston, S.C. (CHS)	32°46'5	79°56'W
Tide Gage No. 3 (CFM)		
Brunswick Intake (BRI)		
Southport (SP)		

Table 3

Radial Distances Between Stations

Stations	Distance
SAV-CHS	130 Km
SAV-WIL	380 Km
SAV-BFT	480 Km
SAV-HAT	620-Km
CHS-WIL	240-Km
CHS-BFT	340-Km
CHS-HAT	480-Km
WIL-BFT	100-Km
WIL-HAT	240-Km
BFT-HAT	140-Km

Savannah and Wilmington airports and Charleston and Cape Hatteras coastal stations. The airport data were undoubtedly contaminated by topographic boundary layer effects and thus may show considerably diminished magnitudes as well as possible rotations in direction relative to the coastal stations. Hourly values of sea level height for 1974-1975 were obtained from the National Survey, NOAA, Rockville, Md. for BFT, FPS and WIL; from an NCSU study sponsored by the N.C. Board of Science and Technology for CHS; and from the UNC Sea Grant College Program for CFM. Three-hourly values of surface wind speed, wind direction, and atmospheric pressure for 1974-1975 were obtained from the National Climatic Center, NOAA, Asheville, NC, for HAT, WIL, CHS, and SAV.

Since we are not interested in motions with frequencies greater than 0.5 cpd, we must average such motions out of the system. We do this by filtering out the high frequency (σ>0.5cpd) motion while allowing the low frequency motion (σ<0.5cpd) to be retained. A Lanczos filter was used to attenuate the daily and semidaily tides and inertial fluctuations inertial periods at/were: SAV/22.64 hr; CHS/22.4 hr; WIL-FPS/21.46 hr; BFT/21.1 hr; and HAT/20.9 hr.). The filter used has its basis in the formula:

$$\frac{\sin(2\Pi(n-1)-(N-1)/(N-1)}{2\Pi(n-1)-(N-1)/(N-1)}$$

where n -0,1,2,3,---N-1: or the number of data samples in the series. The filter envelope is shown in Figure 3. Attenuation at diurnal and higher frequencies was greater than 10^6. After filtering, the sea level data were resampled at 6-hour intervals to create the forty hour low-passed (40HLP) time series.

Three-hourly wind stress vector components were computed from the raw wind speed and directional data. Wind velocity vector components are aligned in a right-handed coordinate system such that east and north are positive. A "northward" wind has the identical meaning as a "southerly" wind, in keeping with accepted convention. Windstress components were computed using a quadratic drag law with the coefficient $C_D = 1.5 \times 10^{-3}$ (Pond, 1975). The wind stress components and atmospheric pressure time series were low pass filtered, using the filter described previously. The atmospheric data were then subsampled at 9-hour

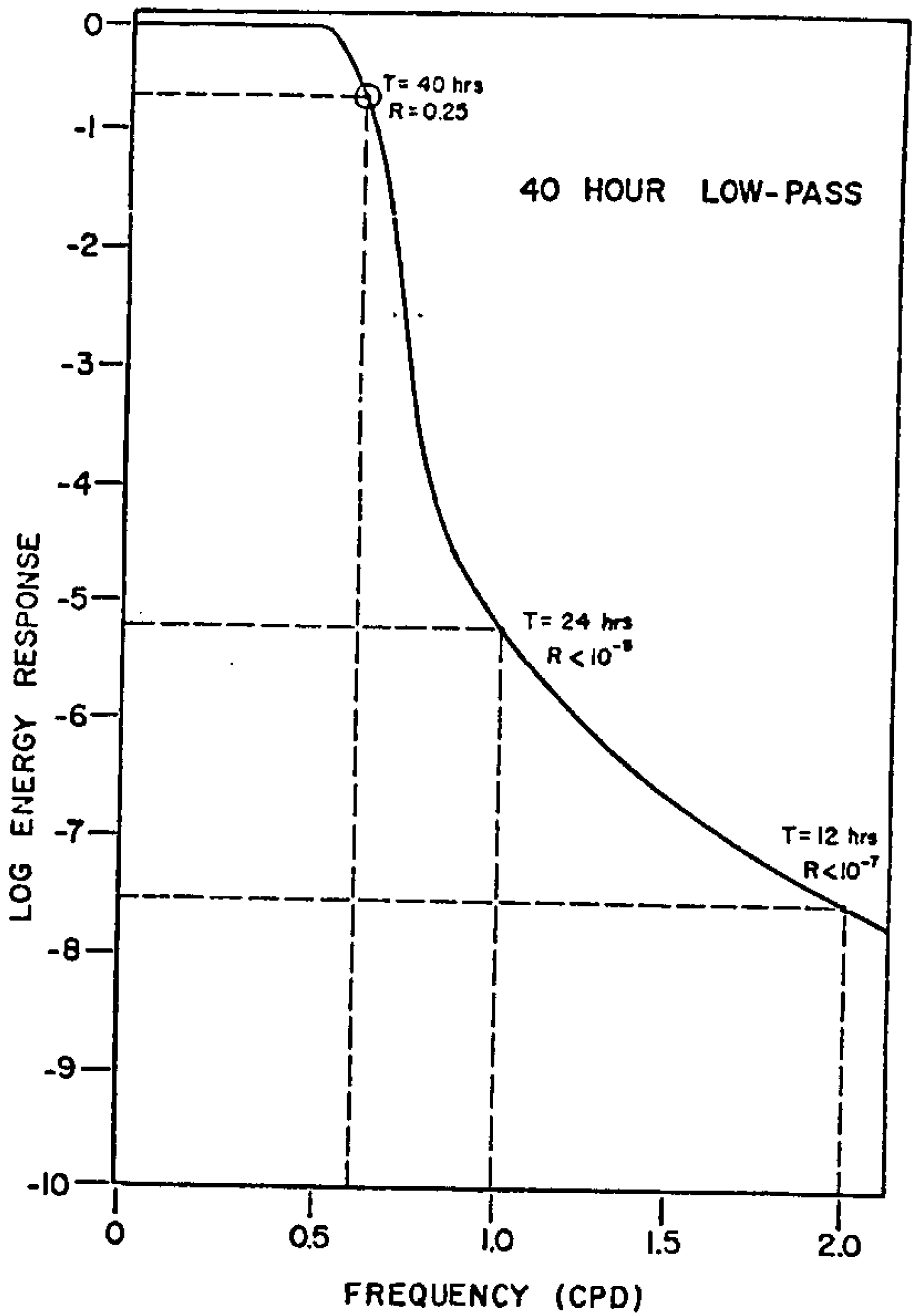


Figure 3: Lanczos cosine taper filter energy response envelope, 40HRLP

intervals and linearly interpolated to 6-hour intervals to be commensurate with the respective sea level data.

Among the suggested factors that significantly affect sea level, particular note is made to the variations in North Atlantic Central Water heat content, atmospheric pressure, atmospheric winds, oceanic currents and continental shelf wave phenomena. The response of the height of the sea surface to varying atmospheric pressure is more correctly expressed in terms of a frequency dependent barometric factor, i.e., a transfer function for the pressure and sea level system. Herein, the sea level data is "adjusted" for the so-called "barometric effect". For frequencies below 1 cpd, the sea surface responds to changes in atmospheric pressure in a reasonably regular, nearly instantaneous fashion. The adjustment suggested over nearly the entire frequency range pertinent herein is approximately 1.05 cm/millibar (eg. Roden, 1960; Mysak and Hamon, 1969), i.e. a one millibar increase (decrease) in pressure depresses (elevates) the sea surface approximately 1 centimeter. Monthly to seasonal to annual variations in sea level, due to the thermal variability of the water column and to changes in the coastal atmospheric climatology, provide appreciable adjustments to a several-day to several-week running average. Examples of the yearly time series of water surface heights and the power spectra of these time series are shown for BFT, CHS and WIL in Figure 4. While these data are presented and described, the seasonal to annual harmonics are eliminated ("demeaned") from the two four-month blocks to be investigated, thereby suppressing the long period trend contamination of the shorter period correlations.

From Figures 4a, b and c, one can see some seasonal to annual changes in the time domain of sea level variability.

During the period January-February, water temperatures are the lowest of the year in the SAB due to atmospheric cooling and sea level is at its yearly minimum, standing 20-25 cm below zero datum. During the period of late winter to early spring, the heat content of North Atlantic Central Water, down to about 100 db, increases and sea level begins to rise, due to density structure alone, not wind effects. This point is well described by Pattullo, et. al. (1955). The high frequency (2 to 10-day) fluctuations superimposed on the seasonal, low frequency background change in sea level are caused by atmospheric cyclones and anticyclones, which spin through the system 5-10 days. Since the SAB is on the southerly side of these storms, the winds observed at the coastal stations generally rotate clockwise as the storm passes by a specific locale. Since sea level variance is strongly tied to wind speed and direction (Pietrafesa et. al., 1978a), water surface heights can rise and fall dramatically over a several-day period.

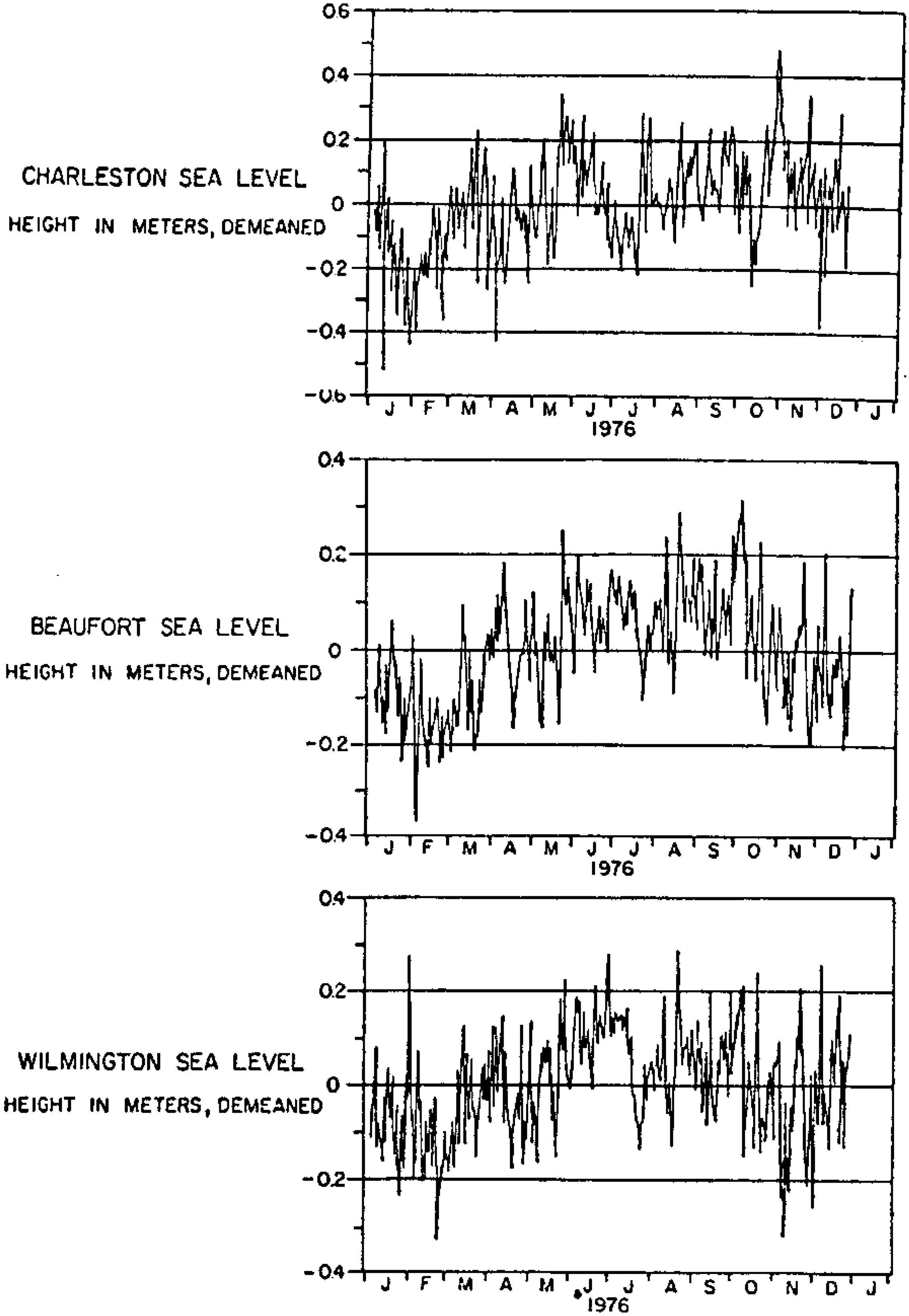


Figure 4: Time series of sea level at Charleston, S.C., Beaufort, N.C., and Wilmington, N.C., 5 Jan. - 27 Dec., 1976

During mid-spring there is a substantial strengthening of the Azores-Bermuda (atmospheric) High, and as the high pressure system expands west and north, a southwesterly wind system develops over the Georgia Bight and migrates northward to the Carolina Capes shelf. The southwesterly winds drive near coastal waters offshore, causing a set-down of sea level at the coast. This sequence of falling sea level occurs in the southerly portion of the SAB and proceeds to the north, with the advent and progression of the southerly to westerly predominant wind fields. When a southwesterly wind blows, sea level at CHS stops its rise as the wind-driven transport of water away from the coast counteracts and eventually overwhelms the rise of sea level due to the increase of North Atlantic Central Water heat content, as described earlier. The response at BFT, with the advent of the southwesterly wind is not as dramatic as that at CHS. This is due to the differences in the topographic geometry of the two sites. While both sites are open coastal stations, CHS is located along a coastline (Figure 2) that changes in a regular, monotonic way in the alongshore direction. BFT, on the other hand, is located at the northwestern corner of a cuspate coastal feature, Onslow Bay (Figure 2), and conventional concepts of wind-driven surface drift do not always hold in such a locale. Current response characteristics for circulation in this region during the summer of 1976 are described in Pietrafesa et. al. (1978b).

Short Term Variations in Sea Level: The Weather

The objective of this portion of the study is to examine sea level fluctuations in the Carolina Capes and the relationships that these fluctuations bear with atmospheric forcing, with coastal geometry, including cape and shoal topography, with local rivers and estuaries and with radiative Gulf Stream effects.

In a study of sea level behavior along the North Carolina coast, Mysak and Hamon (1969) found evidence for the southerly propagation of coherent, subinertial frequency signals. These phenomena were hypothesized to be continental shelf waves, though the authors noted that possible effects of the northerly flowing Gulf Stream and of the local wind field on sea level were not included in their analysis.

Solutions of models of free and forced continental shelf waves in the presence of a sheared current were studied by Brooks and Mooers (1977) and McKee (1977), respectively. These models were then utilized by Brooks (1978) in a study of sea level fluctuations and their relationship to atmospheric forcing at Beaufort and Wilmington. At Wilmington, Brooks found a persistent phase lag 2 to 4 times smaller than the model predicted value; he interpreted this phase lag as evidence of southerly propagating shelf waves. Brooks intimated that the smallness of the phase lag was due to interactions between locally forced and free wave modes. However, since the WIL tide gauge is located 45 km up the Cape Fear River, it is more likely that the observed phase lag is due to an estuarine-induced delay. A significant result of the Brooks (1978) analysis was a large peak in coherency in the cross spectrum at the 3-day period.

As was shown by Pietrafesa et. al. (1978a), this peak in coherency is obvious in other sea level station pairings in the SAB. The significance of this peak is that it coincides with the first mode shelf wave cut-off frequency for the Cape Fear locale, as calculated by Brooks. A similar result was found by Cutchin and Smith (1973) for the Oregon coast. As shown by Meyer (1971), cut-off frequencies are points of potential resonance and thus, if fluctuations in sea level could be shown to be selectively forced at cut-off frequencies, this information would offer credence to the theories suggesting the possible existence of continental shelf waves in the Carolina Capes shelf region between Charleston and Cape Hatteras. It has been shown by Beardsley and Csanady (1979) that north of Cape Hatteras two kinds of pressure fields affect continental shelf waters. The first kind is related to the variability of atmospheric wind forcing and local topography and is trapped within 30 km of the coast by the combined effects of friction, the Coriolis force and bottom topography. The second type of pressure field is an extension of that associated with deep water oceanic gyres, which thus serve as remote forcing agents to the Middle Atlantic Bight (MAB). The disparity in schools of thought concerning the disposition of sea levels in the MAB versus the SAB can be attributed to the fact that in the SAB the Gulf Stream is typically located within 80 -150 km of the continent, while in the MAB its location is more remote from the shelf proper.

Efforts will be made to compare these interpretations with those made in previous investigations, when appropriate. Our results suggest that a majority of the total variance in sea level can be explained as a direct, mechanical response to local wind forcing . Wind components aligned with the local topography were found in general to have more effect on sea level changes than did cross-isobath components. In cases where the cross-shelf component of the wind became important locally, a temporal retardation of sea level set-up and set-down was observed.

The field observations and computational model results presented within indicate that inner shelf waters are dominated by the local wind, the effect of the earth's rotation, topographic frictional effects and bottom slope effects. The e-folding scale of the coupled dynamics is shown to be of the order of 40 km. The only evidence of nonlocal forcing exists in the 12-to 15-day period band, which can be associated with large scale dynamics.

Spectra of subtidal sea level fluctuations were computed for the 4-month period with an 80-hour lag window with 15.7 degrees of freedom. Power spectra of sea level at the three open ocean sea level stations, BFT, FPS and CHS, were similar and had a barely detectable peak at a period of 3.6 days (Figures 5 and 6). The energy level at CHS was higher than at BFT and at WIL for periods longer than 20 days. At the offshore station, i.e. FPS, the energy level was considerably higher than at the one riverine and two coastal stations. All of these power spectra are shown in Figure 6.

Atmospheric pressure was coherent over the whole Carolina coast in 1974 (Pietrafesa et.al., 1978a). However, coastal winds for the year were rather poorly organized (Pietrafesa et.al., 1978a). Wind fluctuations along the Carolina coast were principally in the alongshore direction. Figure 7 shows the alongshore and cross-shelf wind stress spectra at HAT, CHS and WIL. The energy levels of alongshore and cross-shelf wind stress at HAT were considerably higher than those at the southern stations. HAT is the first order National Weather Service Station closest to Beaufort, so the wind stress data there was used in this analysis to represent Beaufort wind stress. The dominant time scale for the alongshore wind stress at WIL and CHS was 7-10 days. However, there was a secondary peak of the alongshore wind spectrum at about a 3.6-day period at both WIL and CHS (Figure 7). A similar secondary peak was observed in the power spectrum of sea levels at the three coastal stations (Figure 6).

The linear correlations of sea level for station pairs are shown in Figure 8. Coherency (γ^2) is generally in excess of 0.6. Exceptions are the coherency between BFT and CHS at periods of 2.5 and 5-7 days. This local minimum of coherence squared is also noticeable between WIL and CHS at 2.5 and 5-6 day periods. Similar minima in the coherency between sea level stations were noted by Beardsley et. al. (1977) and Wang (1979) for the MAB.

Between CHS and BFT there was, in general, a northeastward phase propagation of sea level (Figure 8) consistent with the phase propagation of the alongshore wind stress. A 6-to 12-hour phase lag at WIL with respect to BFT or CHS was observed. The phase lag at WIL was a persistent feature and seemed to be independent of seasons and phase propagation of the alongshore wind and atmospheric pressure (Brooks, 1978; Pietrafesa et. al., 1978a). Pietrafesa (1977) and Brooks (1978) attributed this lag to river effects and continental shelf waves, respectively. In the last section of this report we will present evidence that this lag is associated with the presence of the Cape Fear River Estuary.

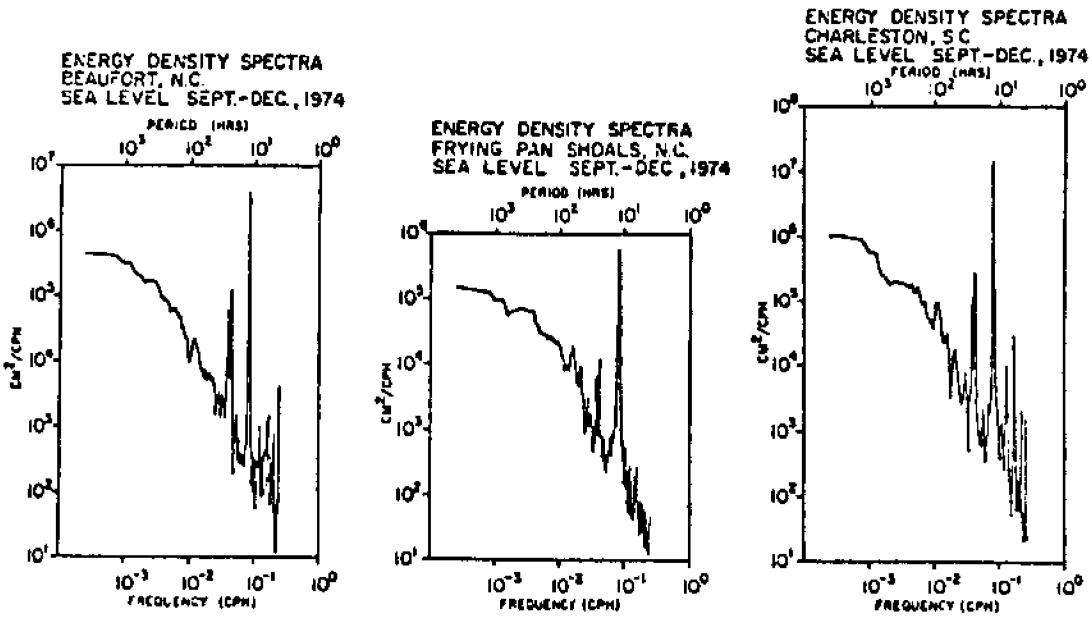


Figure 5: Energy density spectra at Beaufort, N.C., Frying Pan Shoals, N.C., and Charleston, S.C., Sept.-Dec. 1974

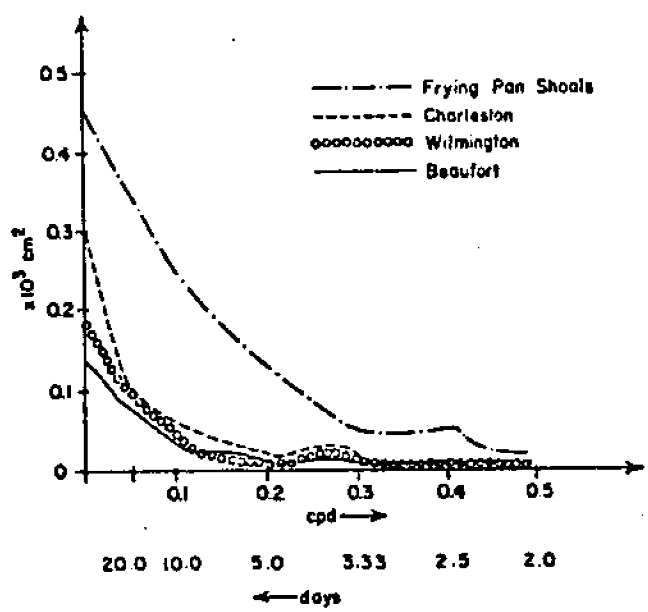


Figure 6: Power spectra of sea level at Frying Pan Shoals, Charleston, Wilmington, and Beaufort, Sept. - Dec., 1974

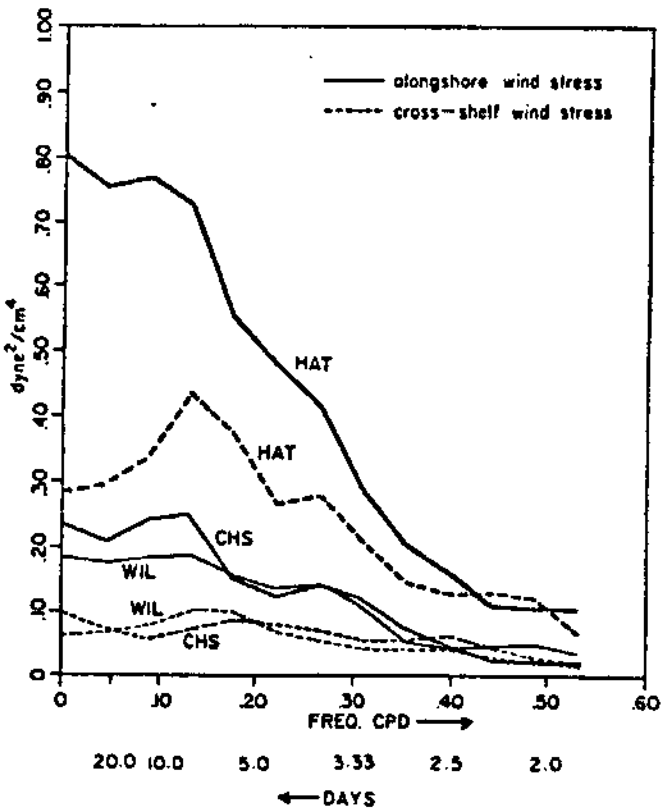


Figure 7: Power spectra of alongshore and cross-shelf wind stress at Cape Hatteras, Wilmington, and Charleston, Sept. - Dec., 1974

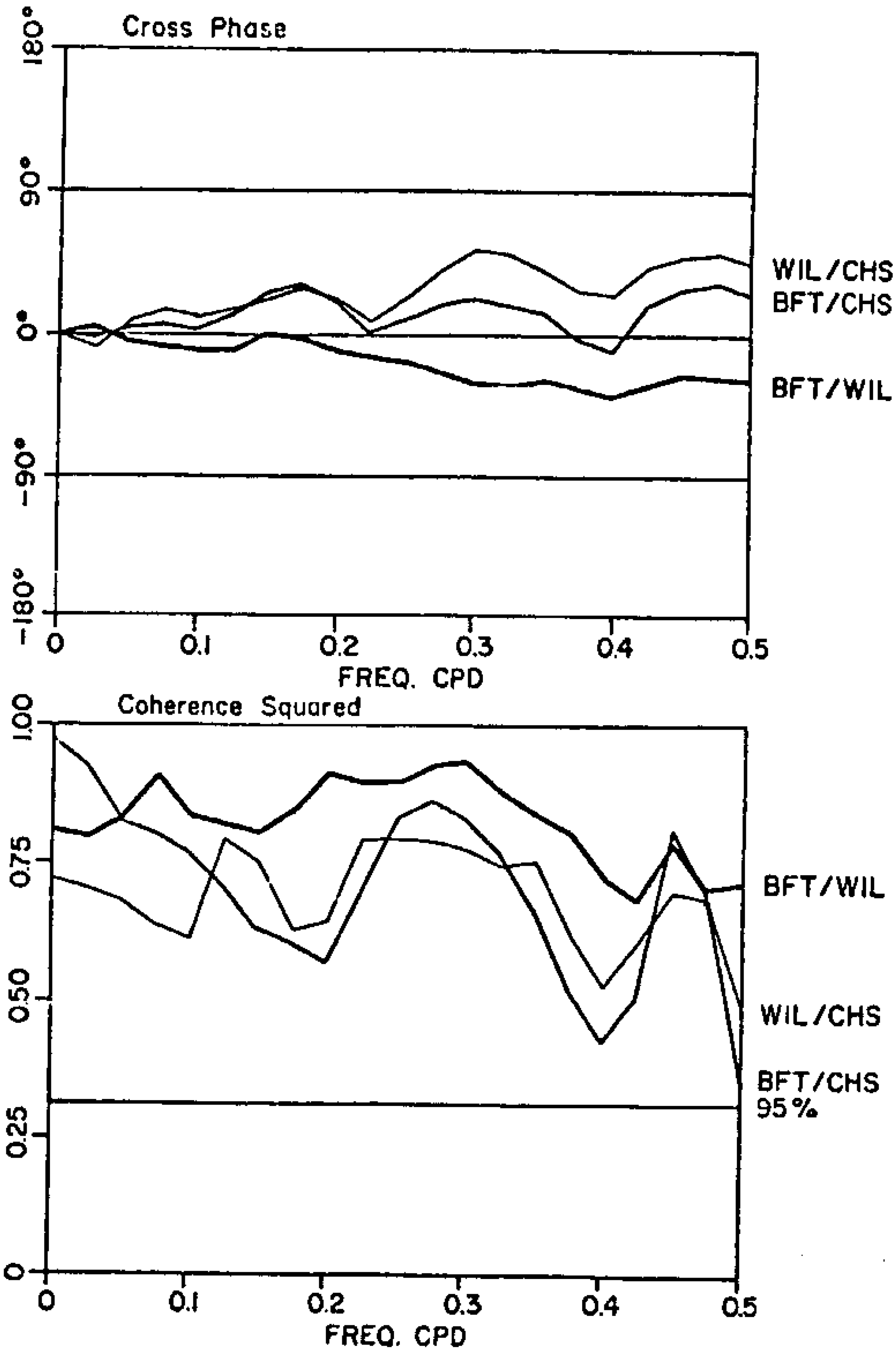


Figure 8: Cross phase and coherence squared of cross spectra of adjusted sea level, Wilmington vs. Charleston, Beaufort vs. Charleston, and Beaufort vs. Wilmington, Sept. - Dec., 1976

Coherences of sea level between FPS and coastal stations were generally low (Figure 9). The consistent phase lag at WIL was not observed at FPS. In spite of the poor correlation, FPS sea level had the highest coherence with CHS. A 52-day period(6 June -27 July, 1975) sea level data set was also available at CFM. The cross spectrum of sea level between CFM and both BFT and CHS were computed over the 52-day segment; the corresponding frequency resolution was 0.06 cpd and the number of degrees of freedom was 12. The number of degrees of freedom for these 52-day computations was different from the previous case and consequently the spectrum results were subject to larger uncertainty. However, as Wang (1979) noted, in the MAB rather constant coherence over different periods usually was observed despite large changes of wind and sea level variance.

The coherence of sea levels between CFM and CHS was high ($\gamma^2 > 0.5$), while corresponding coherence between CFM and BFT was low (Figure 10). This result is consistent with the earlier result (Figure 9) that FPS sea level showed higher coherence with CHS than with BFT. The possible separation of circulation patterns in the Carolina Bays by the extensive excursion of Frying Pan, Lookout and Diamond shoals has been a long standing speculation (Pietrafesa, 1977). Abbe (1895) suggested that "bay eddies" phase-locked with the bays were responsible for the formation of the cuspate coastal boundary. Brooks (1978) also made the same speculation. Mathews and Pashuk (1977) calculated the sea surface circulations from shipboard hydrographic data during February and March 1973 and showed that they indeed were separated by Frying Pan Shoals. Our results indicate that there was definite uncoupling of circulation patterns between Long and Onslow Bays during the 52-day period (6 June - 17 July 1975).

In essence, except for 2.5-day and 5-day fluctuations, coherence of sea levels between coastal station pair was high, suggesting that the three coastal stations were subjected to common forcing. However, coherence was low between FPS and the coastal stations. Csanady (1978) postulated that the inner shelf of the MAB was dominated by friction, while the outershelf was dominated by large scale gyres. In this regard, our results seem to suggest that Ekman set-up/down may dominate in the inner shelf region, and nonlocal forcing, such as direct or radiative effects of the Gulf Stream, may dominate over the outer shelf.

Relations between local wind and sea level were examined from cross spectrum analysis. Coherence was high between along-shore wind stress and sea level at CHS and BFT, the coastal stations (Figure 11). Exceptions were at WIL for periods of 2.5, 5, and 12 to 15 days. The drop of coherence at 2.5-day and 5-day periods also were present at CHS but were not so profound at BFT. This finding was entirely consistent with the fact that the alongshore wind stress was poorly correlated between CHS and WIL but well correlated between WIL and HAT at 2.5 and 5-day periods

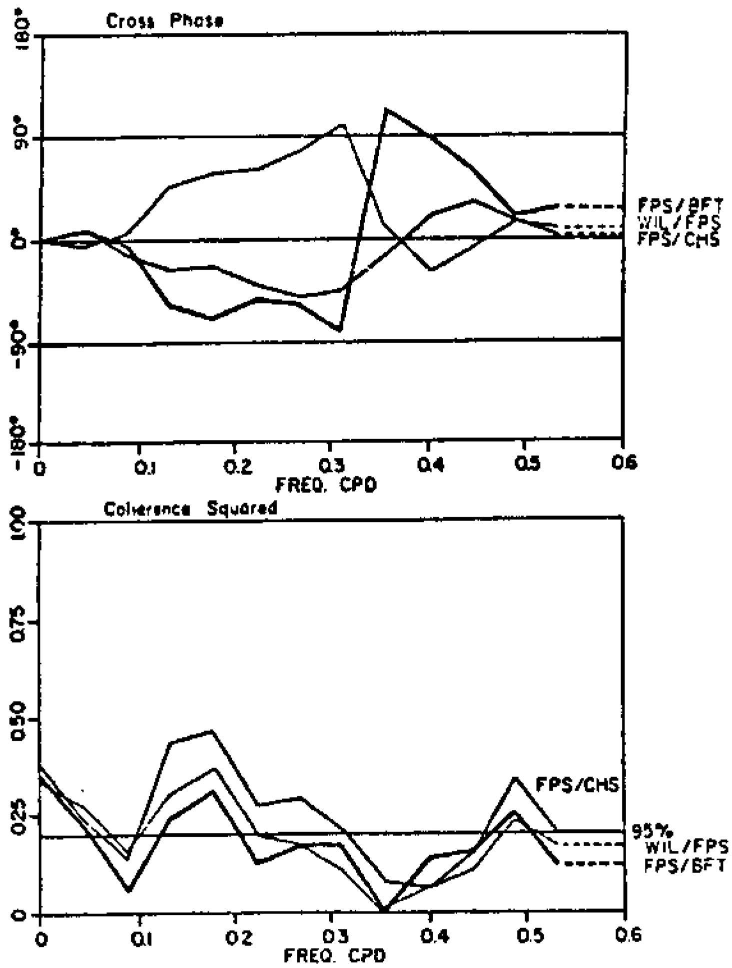


Figure 9 Cross phase and coherence squared of cross spectra of adjusted sea level, Wilmington vs. Frying Pan Shoals, Frying Pan Shoals vs. Charleston, and Frying Pan Shoals vs. Beaufort, Sept. - Dec., 1974

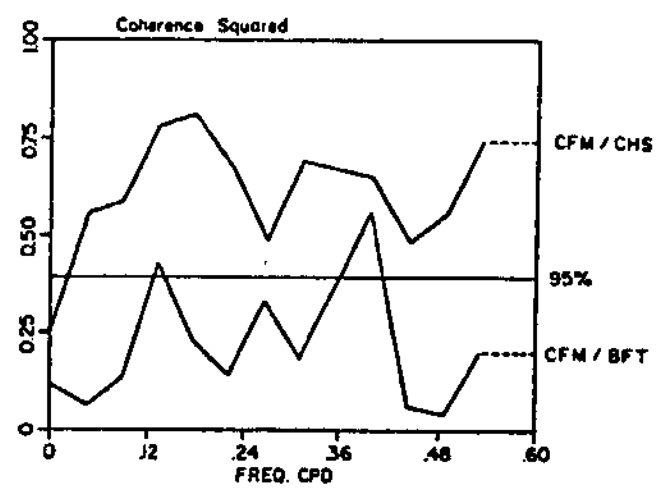


Figure 10: Coherence squared of cross spectra of adjusted sea level, Tide Gauge 3 vs. Charleston, and Tide Gauge 3 vs. Beaufort, 6 June - 27 July, 1975, degree of freedom = 12.0

(Figure 12). The implication was that a significant alongshore fetch was needed for the alongshore wind to set up coastal sea level effectively. Apart from the 12-to 15-day period band, subtidal sea level fluctuations at the three coastal stations could be attributed to direct, local wind forcing. The 12-to 15-day period minimum of coherence of alongshore wind stress versus sea level was evident at all three coastal stations and FPS and apparently was not related to wind stress (Figures 11 and 12.)

The cross phase of alongshore wind vs. sea level in the capes region (Figure 11) was similar to that observed by Wang (1979 at Sandy Hook. He described the fluctuations in sea level there to be locally forced by the wind with an adjustment time of frictional set-up/down of approximately nine hours. The exception in the capes data to this finding occurred at WIL, where the river delay effect retarded the set-up/down for 6 to 12 hours.

At FPS, sea level fluctuations were more responsive to cross-shelf wind than to alongshore wind (Figure 11). Since the local isobaths are in the NW-SE (cross-shelf) direction (Figure 2, FPS sea level appeared to be driven by along-isobath wind stress to a certain degree. The poor correlation between wind stress components and sea level ($\gamma^2 < 0.5$) could be attributed to the speculation that an offshore station was somewhat seaward of the dominant influence of frictional set-up and was influenced predominantly by "nonlocal" forcing, such as the Gulf Stream provides.

A Simple Conceptual Model

A complete model describing the nearshore wind driven circulation of the Carolina coast would include the along-shore geometry, temporally varying wind stress, the Gulf Stream and an irregular coastline, among other factors. However, the model we present below is deliberately simplified in those respects and is directed only at isolating predominant factors which are believed to influence sea level fluctuations in the Carolina Capes region.

Consider a semi-infinite ocean bounded by a straight coastline (along the y-axis), with the x-axis pointing eastward offshore and the y-axis positive to the north in a right-handed system. The vertically averaged equations of motion and continuity are:

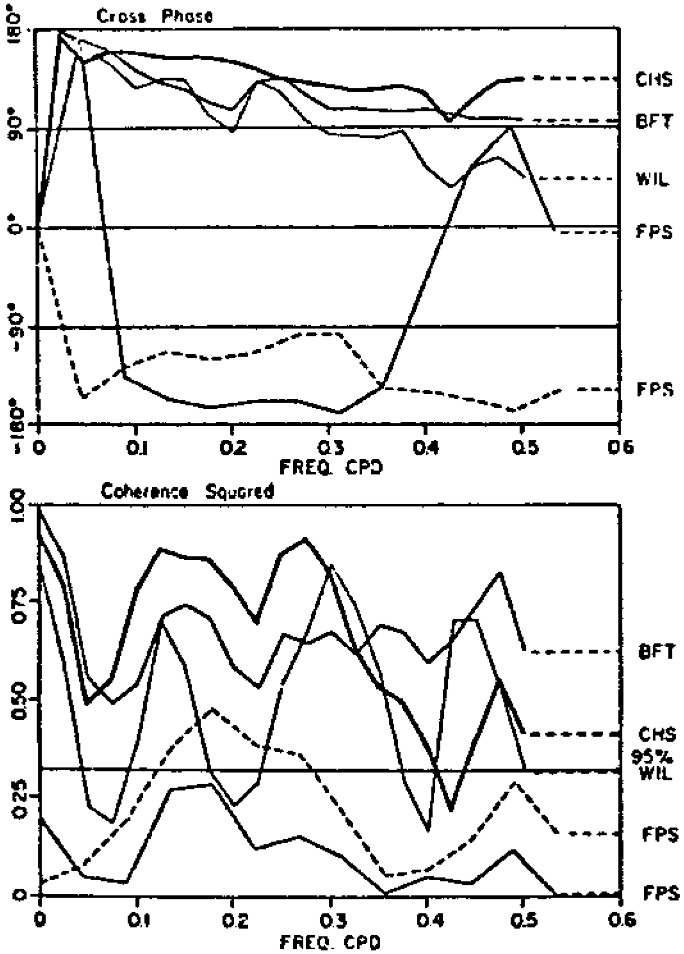


Figure 11: Cross phase and coherence squared of cross spectra of alongshore wind stress (——) or cross-shelf wind stress (---) vs. adjusted sea level at Charleston, Beaufort, Wilmington, and Frying Pan Shoals, Sept. - Dec., 1974

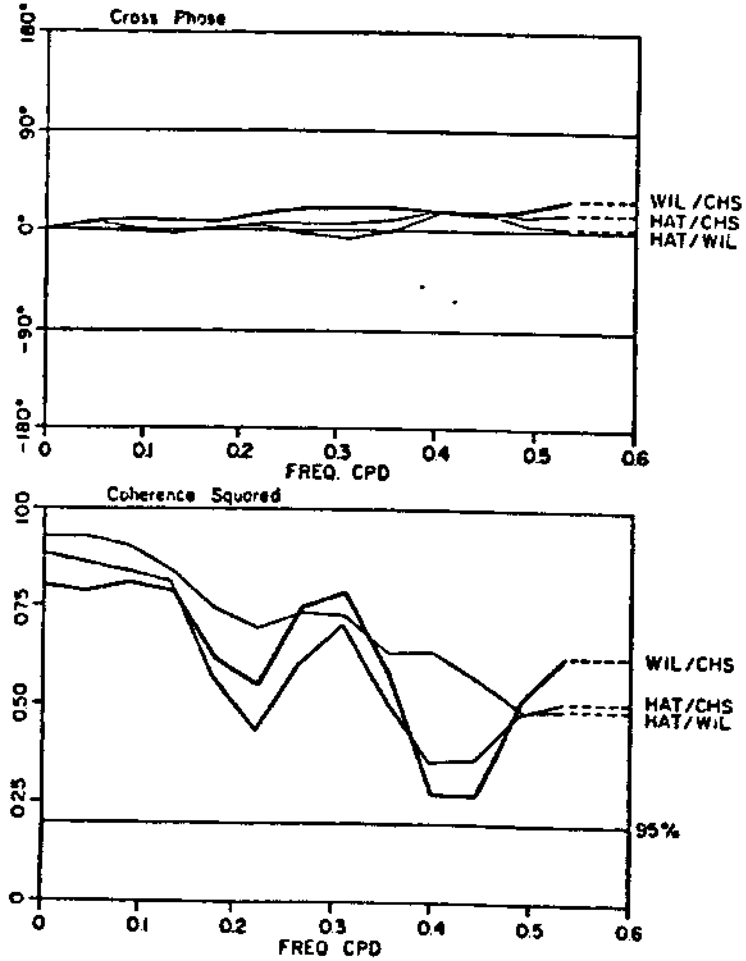


Figure 12: Cross phase and coherence squared of cross spectra of alongshore wind stress, Wilmington vs. Charleston, Wilmington vs. Cape Hatteras, and Charleston vs. Cape Hatteras, Sept. - Dec., 1974

$$\frac{\partial u}{\partial t} - fv = g\zeta_x + \frac{X}{h} - \frac{ru}{h} \qquad (1a)$$

$$\frac{\partial v}{\partial t} + fu = \frac{Y}{h} - \frac{rv}{h} \qquad (1b)$$

and

$$\frac{\partial}{\partial x}(hu) = \frac{\partial \zeta}{\partial t} \qquad (1c)$$

where ζ is the sea level elevation, Y is the cross-shelf and alongshore components of wind stress, h(x) is the equilibrium water depth, and r is a bottom frictional coefficient with the dimensions of velocity. We also assume that the wind stress and bottom topography are uniform in the alongshore direction so that only localized response are investigated. We impose a monochromatic, clockwise rotating wind field of the form

$$X = -ia\, e^{i\omega\tau}$$
$$Y = e^{i\omega\tau} \qquad (2)$$

where $0 < a < 1$ is a measure of the importance of the cross-shelf wind stress with respect to that of the alongshore wind stress. The imposed wind field (2) resembles actually observed wind fields (Pietrafesa, et. al., 1978a). A governing equation for u can be obtained by combining relations (1) and (2) so that

$$u_{xx} + 2\frac{hx}{h}u_x + \left[\frac{\frac{2r}{h}\omega^2 - \frac{i\omega r^2}{h} + i\omega^3 - i\omega f^2}{g(r + i\omega h)} + \frac{h_{xx}}{h}\right]u \qquad (3)$$

$$= -\left[\frac{X_{tt} + \frac{r}{h}X_t + fY_t}{gh(r + i\omega h)}\right]$$

with boundary conditions

$$U = 0 \text{ at } x = 0,\ x \rightarrow \infty . \tag{4}$$

Equation (3) then is applied to the cross-shelf section at CHS, where the nearshore bathymetry is relatively uniform in the alongshore direction. Figure 13 depicts the cross-shelf bottom topography at CHS as a solid line, while the dashed line is a fitted curve obeying the exponential profile

$$h(x = 10 + e^{5.05(10^{-5})x} \quad \text{(in meters)}. \tag{5}$$

The fitted curve agrees well with the CHS cross-shelf bottom profile from the coastline to 100 km offshore, roughly the location of the inshore edge of the Gulf Stream. A general idea of the nature of the sea level response may be obtained by considering the sloping shelf limit under which

$$\frac{h_{xx}}{h} \gg \frac{\dfrac{2r\omega^2}{h} - \dfrac{i\omega r^2}{h} + i\omega^3 - i\omega f^2}{g(r + i\omega h)} . \tag{6}$$

The solution for ζ can then be obtained from (1c) and (3) and can be written as

$$\zeta(x) = -\frac{i}{g}\int_x^\infty \left(\frac{a}{h} + \frac{f\omega h}{r^2 + \omega^2 h^2}\right) dx + i\int_x^\infty \frac{fr}{\omega^2 h^2 + r^2}\, dx . \tag{7}$$

One can see from equation (5) that both integrands are positive definite functions; therefore, $\zeta(x)$ is a monotonic decreasing function of x. The cross phase of the alongshore wind versus the coastal sea level can be expressed as

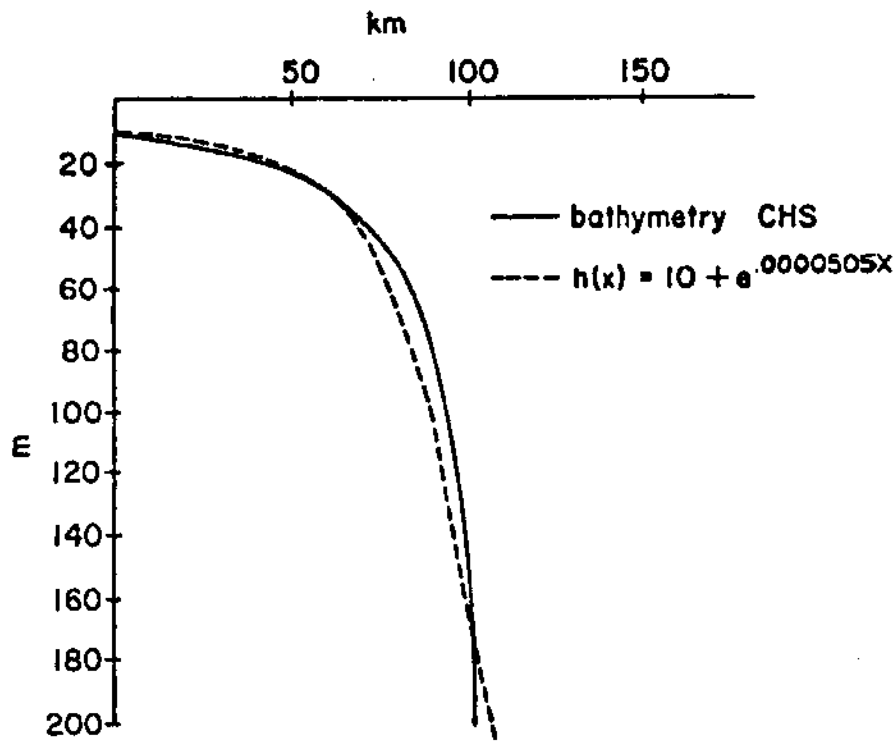


Figure 13: Cross shelf bottom topography at Charleston (——) and exponentially approximated bottom law (---)

$$\Theta = 90° + \tan^{-1} \frac{\int_0^\infty \frac{fr}{\omega^2 h^2 + r^2} dx}{\int_0^\infty \left(\frac{a}{h} + \frac{f h \omega}{\omega r^2 + \omega^2 h^2}\right) dx} , \qquad (8)$$

which is a decreasing function of ω. As $\omega \to 0$, Θ approaches 180°. Thus $\Theta(\omega)$ resembles the observed cross phase (Figure 11) qualitatively.

The effect of Gulf Stream fluctuations on coastal sea level could be estimated from equation (3), but for our purposes we will assume an "effect of Gulf Stream forcing" by using the offshore boundary constraint

$$u = U_o e^{i\omega t} \quad \text{at } x = 100 \text{ km} \quad . \qquad (9)$$

In the absence of wind forcing, it follows that

$$(hu)_x = \text{constant} \quad ; \qquad (10)$$

therefore, ζ is constant in x. This is a straightforward consequence of mass conservation and provides only an upper bound for the coastal sea level change due to Gulf Stream fluctuations. In reality, the oscillation of the Gulf Stream boundary which is not an oscillating wall, attenuates shoreward without alongshore uniformity. Observational evidence (Webster, 1961; Hansen and Maul, 1970; Pietrafesa, 1978; Legeckis, 1979) indicates that Gulf Stream advected frontal waves of wavelength the order of 150-200 km are a dominant feature in the Carolina Capes region. Additionally, theoretical computations (Orlanski, 1969; Niiler and Mysak, 1971; Rooney, et. al., 1978; and Chao and Janowitz, 1979) indicate sound physical bases for both stable and unstable wave modes existing and being advected along the Gulf Stream Front between CHS and HAT. These waves are suspected, from both observational license and theoretical insights, to decay shoreward with e-folding scales of the order of 30 km.

To compute the complete solution of relation (1), the governing equation for ζ may be derived as

$$\zeta_{xx} + \left(\frac{2rH + 3i\omega h_x}{r + i\omega h} - \frac{2((f^2-\omega^2)hh_x + i\omega rh_x}{(r^2 - \omega^2h^2 + f^2h^2) + 2i\omega rh)} \right) \zeta_x -$$

$$\frac{i\omega((r^2 - \omega^2h^2 + f^2h^2) + 2i\omega rh)}{gh^2 \ (r + i\omega h)}$$

$$= \left(\frac{rH + 2i\omega h_x}{gh(r-i\omega h)} - \frac{2((f^2 - \omega^2)h_x + i\omega rH)}{g((r^2 - \omega^2h^2 + f^2h^2) + 2i\omega rh)} \right) X \qquad (11)$$

$$+ \left(\frac{2fH}{g(r + i\omega h} - \frac{2f(f^2 - \omega^2)hh_x + i\omega rh_x)}{g(r + i\omega h) \ ((r^2 - \omega^2h^2 + f2h2) + 2i\omega rh)} \right) Y ,$$

where $H \equiv h_x/h$. The boundary conditions for ζ equivalent to (4) are

$$\zeta_x = \frac{fY}{g(r + i\omega h)} + \frac{X}{gh} \quad , \text{ at } x = 0 \qquad (12)$$

$$\zeta = 0 \qquad , \text{ at } x \to \infty \quad .$$

Equations (11) and (12) with specified wind stress (2) and topography (5) are now solved utilizing the numerical scheme of Lindzen and Kuo (1969).

Figures 14a and 14b show the amplitude of sea level and cross phase of alongshore wind versus sea level, respectively, as a function of dimensionless frequency (ω/f) for a range of frictional coefficients (r). The alongshore wind stress is taken to be 1 dyne/cm^2, a is 0.1, and the Coriolis parameter is f -.89 x 10^{-4} sec^{-1}. An augmented frictional effect tends to hamper coastal sea level set-up (Figure 14a). In the limite of $\omega \to 0$,

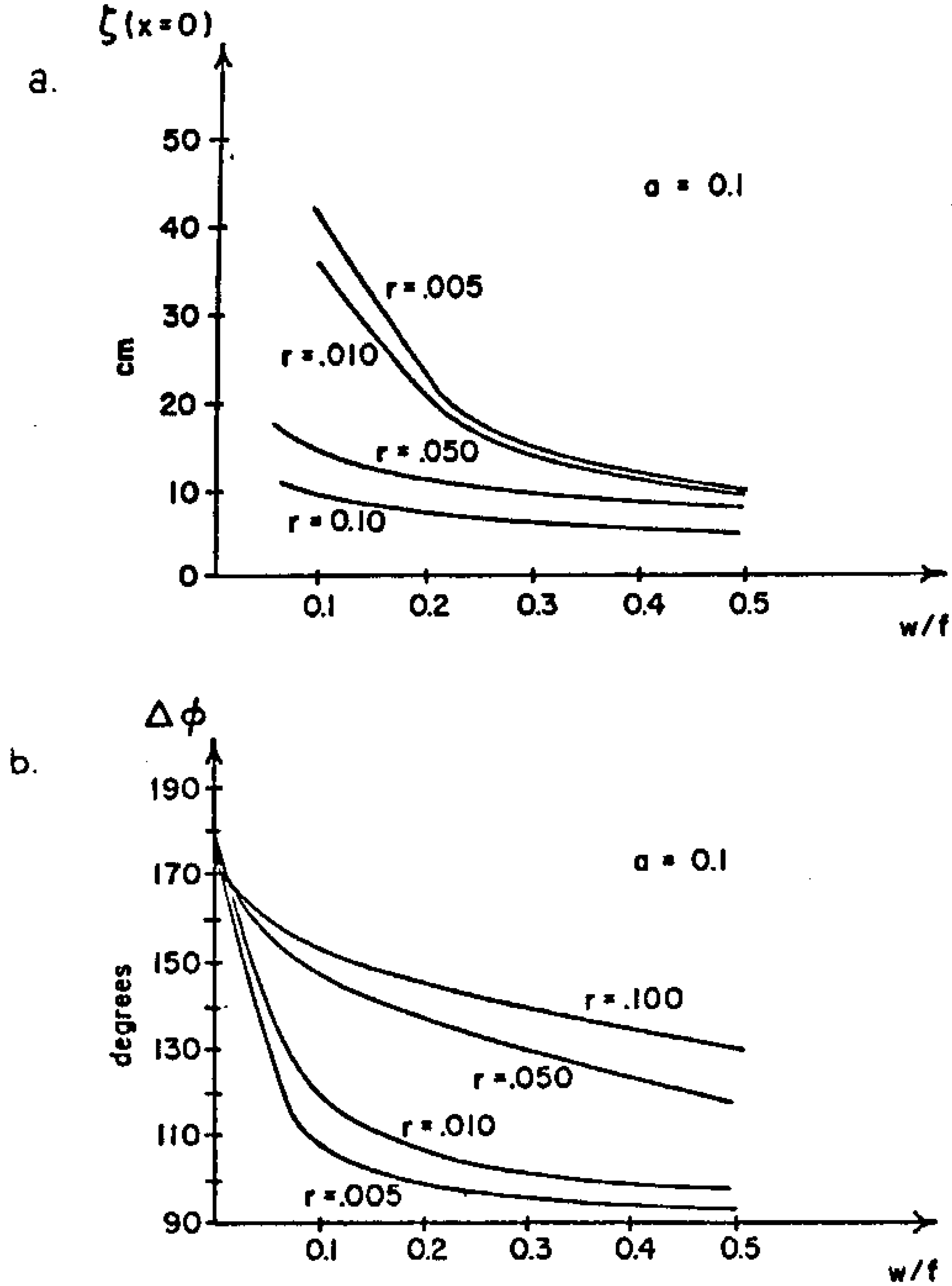


Figure 14: Model-predicted a) amplitude of sea level and b) cross phase of alongshore wind vs. sea level.

the solution for ζ blows up at $x = 0$. Alongshore variation of wind stress must be incorporated to make the problem mathematically wellposed, and the solution of arrested topographic Rossby waves (Csanady, 1978) is recovered. Csanady showed his wave to be arrested essentially in space (alongshore scale) while our wave is arrested in time. Larger friction also is shown to hasten the response of sea level to alongshore wind (Figure 14B). In the frictionless limit, alongshore wind and sea level are in quadrature at all frequencies. In a comparison of observed cross phase (Figure 7) with model results, it appears that 4 = 0.05 cm/sec is a reasonable estimate for the frictional constant over the Carolina Capes shelf.

Figure 15 shows the sea level amplitude versus offshore distance for a = 0.1 and r = 0.05 cm/sec. At all frequencies the sea level decreases montonically offshore as predicted in equation (7). The set-up/down favors a low frequency wind event. The e-folding scale of set-up is in general less than 40 km. Thus for an intense boundary current located more than 100 km offshore, the cross-shelf fetch of the wind stress responsible for coastal sea level set-up is less than 100 km. Thus, the amplitude of coastal sea level set-up may be computed as if the Gulf Stream were absent. This notion is further supported by by observational evidence (Figure 17), as will be shown.

Figure 16 shows the cross phase of alongshore wind versus ω/f for 4 -.05 cm/sec and several values of a. It should be noted here that a is a measure not only of the actual cross-shelf wind stress but also of its actual effectiveness. For example, the local isobaths at Cape Fear favor the cross-shelf wind, and the Cape Fear River is aligned in a somewhat cross-shelf direction. The value of a WIL should be larger than at other coastal stations. This is justified from the observation that the correlation between the cross-shelf wind and sea level at FPS (Figure 11) and WIL (results not shown) is higher than other coastal stations. The cross phase of Y versus ζ decreases with increasing a. Apart from any river delay, part of the time lag at WIL (Figure 11) could be attributed to this effect. The effect of Gulf Stream fluctuations on coastal sea level also could be estimated. The computed result essentially confirms the argument of the sloping shelf limit, hence is not presented here.

Discussion of Results

In the Carolina Capes shelf region, coastal sea level response to wind forcing has considerable variation. Within the 2-to 10-day period band, sea level fluctuations can be attributed to local wind forcing (Figure 11). There is also some evidence

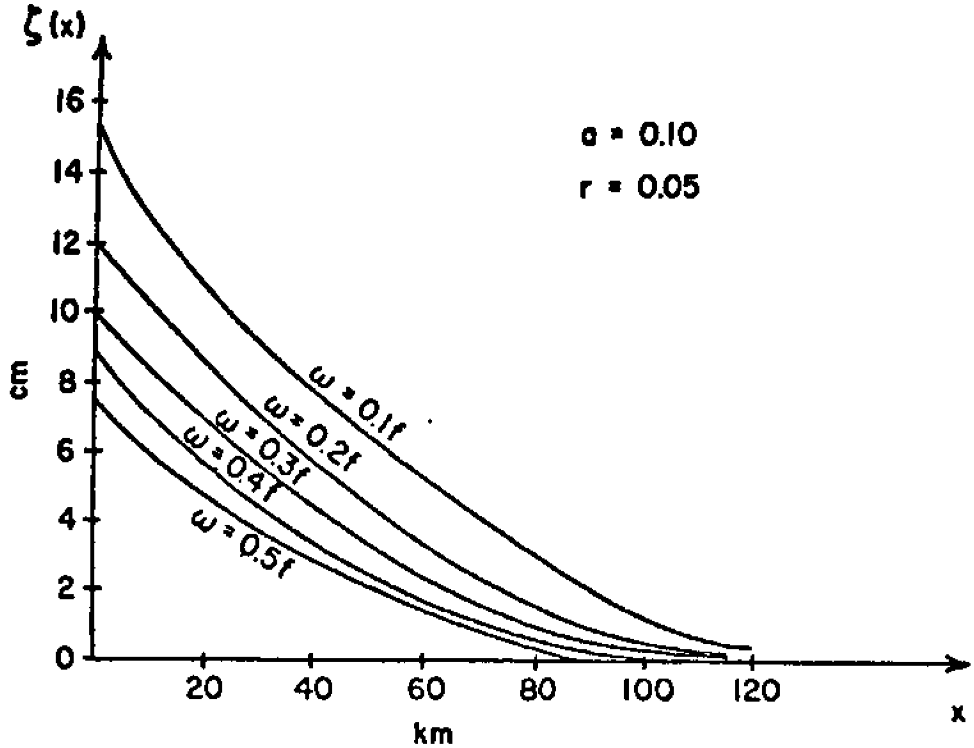


Figure 15: Model-predicted sea level amplitude vs. offshore distance for a = 0.1 and r = 0.05 cm/sec as a function of dimensionless frequency (/f)

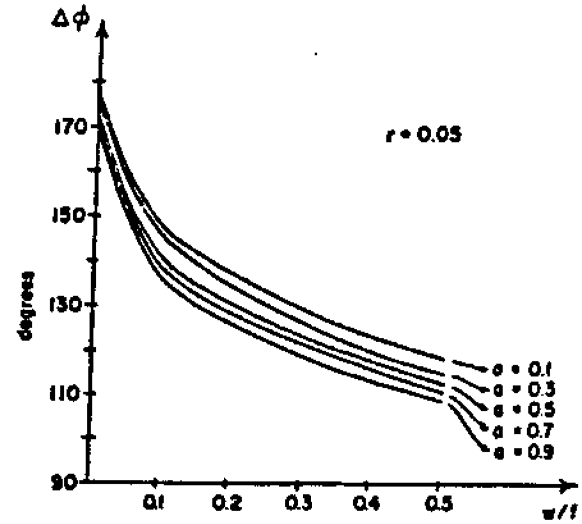


Figure 16: Cross phase of alongshore wind vs. dimensionless frequency for r = 0.05 cm/sec and varying a

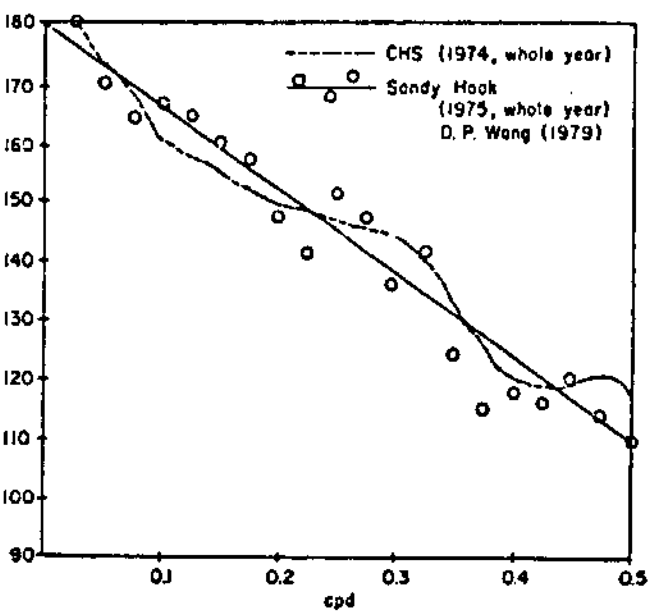


Figure 17: Cross phase of alongshore wind stress vs. adjusted sea level at Charleston, S.C., for 1974 and at Sandy Hook, N.J., for 1975 (D.P. Wang, 1979)

that the subtidal wind-driven circulation nearshore is separated from Long Bay to Onslow Bay by the large excursion of shoals off Cape Fear (Figure 9). Along-isobath wind forcing is found to be more important than is cross-isobath wind forcing.
A larger contribution from cross-shelf wind forcing usually results in a phase lag of coastal sea level fluctuation response (Figure 16).

Sea level fluctuations of periods of 12-15 days could not be attributed to local wind forcing and likely were caused by instabilities of the Gulf Stream or large scale phenomena. Brooks (1978) attributed these fluctuations to atmospherically forced disturbances. The moderate correlation of long period sea level fluctuations with atmospheric pressure could be visualized even in the time domain, which led Brooks (1978) to suggest a reasonance mechanism at appropriate cut-off frequencies. Attractive as it was, the question as to why "resonance" was forced by local atmospheric pressure instead of alongshore wind stress was unanswered. The other possibility was that the 12-15-day periods of nonlocal forcing were in fact caused by large scale circulation, which in turn correlated well with atmospheric pressure. More observational work is definitely needed to resolve this issue.

In the presence of an intense sheared current, local Ekman suction will be modified by the current shear as well as the Ekman choking effect (Mooers, 1977). This fact can be used to add further testimony to questions concerning the importance of the Gulf Stream to coastal sea level set-up. Figure 13 shows the cross phase of alongshore wind stress versus adjusted sea level at CHS during 1974 and at Sandy Hook during 1975, as obtained by Wang (1979). The Gulf Stream is ~ 500 km offshore of this site. One can see from Figure 17 that the presence of the Gulf Stream, approximately 100 km offshore from CHS, poses no fundamental change to the supposition that MAB concepts of sea level response to wind forcing also apply in the SAB. The time lag for alongshore wind set-up of sea level is a constant of eight hours. The implications are that the cross-shelf fetch responsible for coastal sea level set-up is less than 100 km and that coastal sea level is to a large extent subjected to local wind forcing.

The e-folding scale of sea level set-up is in general less 40 km. Thus at FPS (70 km offshore), sea level set-up is roughly 1/5th of the coastal set-up. As a consequence, the coherence between sea level at FPS and the coastal stations (Figure 9) is low, as is Y versus ζ at FPS (Figure 11). In contrast to sea level set-up, the energy level of sea level at the offshore stations is considerably higher than that of the coastal stations (Figure 2). In the light of vorticity dynamics (Buchwald and Adams, 1968), this may be an indication that Gulf Stream frontal oscillations decay, i.e. are attenuated rapidly, in the shoreward direction.

oscillations decay, i.e. are attenuated rapidly, in the shoreward direction.

Shelf Waves: Do They Exist in the Carolina Capes?

As previously noted, sea level data collected at several coastal stations have been used to "confirm" the speculated existence of barotropic continental shelf waves propagating southerly along the coastline of the southeastern U.S. The region under consideration here is the coastal margin between Beaufort and the mouth of the Cape Fear River Estuary (Figure 2). Mysak and Hamon (1969) first suggested that rightbounded along-shelf waves were present in sea level data collected at Southport, N.C., near the mouth of the Cape Fear, and at Beaufort (Figure 18) during the 1953-1954 period. The conclusions reached by these authors is not questioned. More recently, however, Brooks (1978) concluded from a study of Beaufort and Wilmington (Figure 2) tide gauge data that the adjusted sea level phase lags between Beaufort (BSL) and Wilmingtion (WSL) indicated the presence of the first three zero group velocity barotropic shelf wave modes. It is of note, however, that the station to station phase lags computed by Brooks for the three wave modes were of the order of a quarter to a half of those predicted from conventional shelf wave theory (LeBlond and Mysak, 1978). Additional information provided herein shows that not only does WSL lag BSL but that WSL also lags sea level at the mouth of the Cape Fear River (MSL). While WSL always lags BSL, MSL and those subtidal data collected 62.5 km seaward of the estuary mouth at Frying Pan Shoals (FSL) actually lead BSL in the two-day to two-week period bands, which are characteristic of shelf waves (LeBlond and Mysak, 1978).

The sea level phase propagation evidence to be presented herein suggests that it is more probable that the WSL phase lags relative to other coastal stations referred to are best explained in terms of an estuarine-river associated lag and not in terms of southerly propagating barotropic shelf waves. The Cape Fear River Estuary is a tidally and atmospherically dominated, partially stratified coastal plain estuary (Dyer, 1973). The WIL tide gauge is located 45 km from the mouth of the estuary; the Brunswick Intake gauge (BRI) is located 11 km from the mouth and CFM is located at the mouth (Figure 18)

Figures 19a, b, c, and d (19e and f), show coherency and phase for WSL, MSL, BRI Sea level (ISL) and BSL (FSL, BSL and WSL) station pairs during the period 2 June -1 August 1975 (1 September -31 December, 1974). One can see from Figures 19a, b, c and f that WSL lags BSL, MSL, ISL and FSL in all frequency bands. This result alone is not sufficient proof of the absence of shelf waves. However, the cross correlations between MSL and

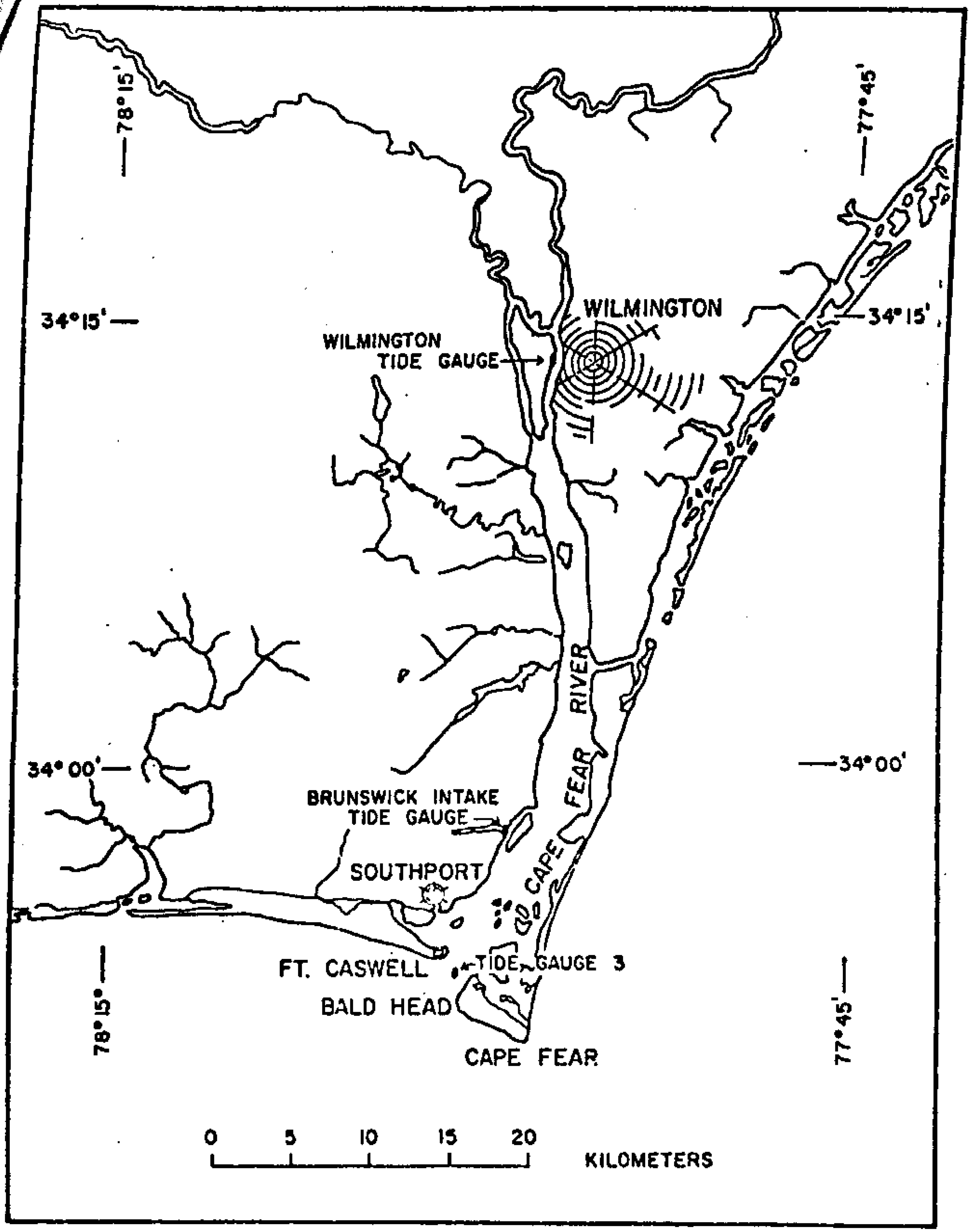


Figure 18: Tide gauge locations along the Cape Fear River extending from the mouth of the estuary to Wilmington

BSL (Figure 19d) and FSL and BSL (Figure 19e) indicate that BSL lags MSL and FSL for all fluctuations having periods greater than 2.55 days and coherence spectral peaks above the 95% confidence level. This is indicative that WSL lags all stations (BSL, MSL, FSL, and ISL), event though MSL and FSL lead BSL to the north. This scenario is depicted in Figure 20.

The phase lag of WSL relative to MSL could be computed as half of the period of the fundamental mode of a seiche in a narrow channel (Proudman, 1952) or rather the period for a gravity wave to propagate upstream. For the Cape Fear, this period is of the order of 1 1/4 hours. Yet the phase lag of WSL behind MSL is of the order of 5-6 hours, which suggests that perhaps stratification or river discharge effects oppose the upstream propagation of events and the net result is an apparent delay of phase. This is conjectural at this time, but the point remains that while WSL lags all other coastal stations, MSL and FSL lead BSL, which leads to the conclusion that coherent sea level phenomena either propagate or are advected northward along the North Carolina coast in Onslow Bay by the Gulf Stream.

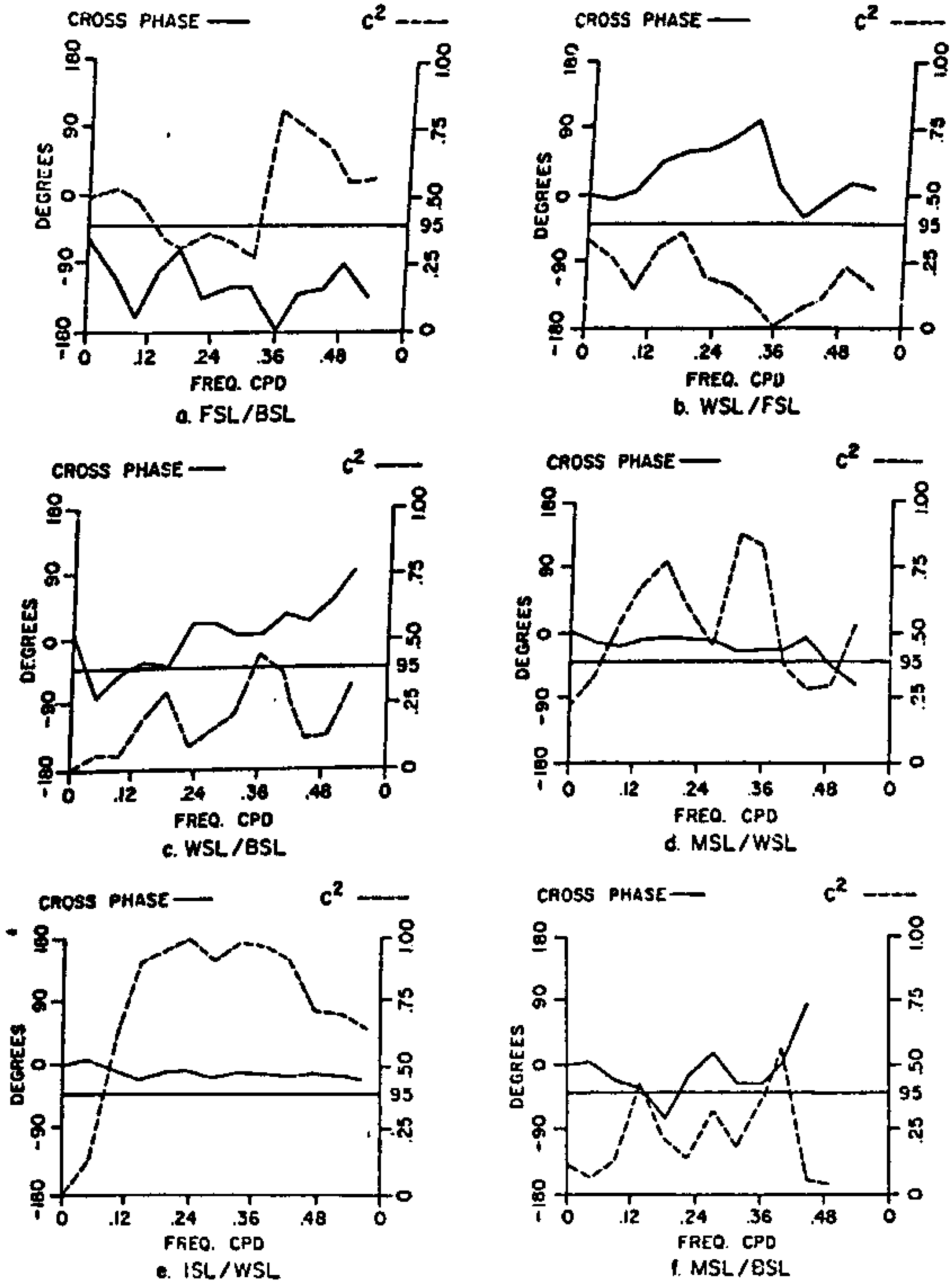
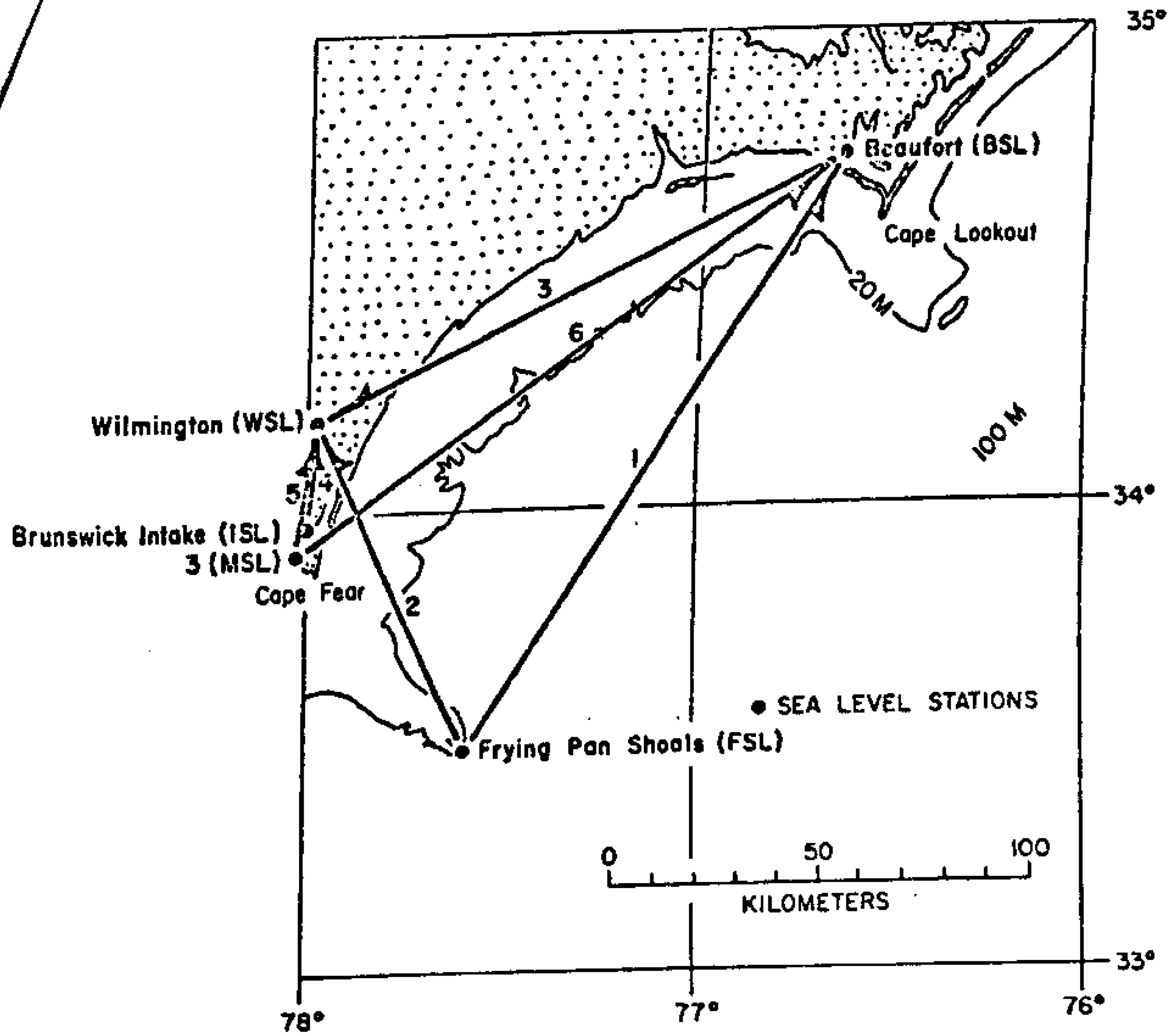


Figure 19: Coherence squared and phase between sea level time series with 95% significance level denoted by the number 95 for a) Frying Pan Shoals and Beaufort, b) Wilmington and Frying Pan Shoals, c) Wilmington and Beaufort, d) Tide Gauge 3 and Wilmington, e) Brunswick Intake and Wilmington, and f) Tide Gauge 3 and Beaufort.



(1) Figure 19a
(2) Figure 19b
(3) Figure 19c
(4) Figure 19d
(5) Figure 19e
(6) Figure 19f

Figure 20: Sea level phase lag where phase at tail of arrow leads phase at head, i.e. x leads y for (x) ⟶ (y)

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