WHOI-84-14

**Analysis of Flow Within the Coastal Boundary
Layer Off Long Island, New York**

by

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April 1984

Technical Report

*Funding was provided by the Department of Energy
under contract DE-AC02-79EV10005.*

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Woods Hole Oceanog. Inst. Tech. Rept. WHOI-84-14.*

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Abstract

From 1974 through 1978 a series of intensive measurements were made in the coastal waters within 12 km of Long Island. The data were derived from two sources: a mooring array from which time series of temperature, salinity and water velocity were measured at four depths at each of four offshore distances; and high resolution, daily hydrographic surveys. Analysis of subtidal cross-shore velocity fluctuations has indicated a two-layer response to wind forcing, with near-surface flow to the right of the longshore wind and opposing flow below. The magnitude of these fluctuations increased in the seaward direction on a scale nearly equal to the internal deformation radius. The phase between longshore velocity fluctuations and longshore wind stress approached zero with decreasing bottom depth, probably the result of bottom stress. The vertical structure of longshore fluctuations during stratified conditions markedly differed from that during unstratified conditions, and resembled the structure derived from a simple two-layer coastal flow model. Significant mean offshore flow was measured during experiments in August and September, despite negligible mean wind stress during the same periods. This flow was most likely due to persistent longshore density gradients, as are consistently inferred from hydrographic data taken in the vicinity.

1. Introduction

Physical oceanography of the coastal ocean has not received significant attention until as recently as 15 years ago. Despite this rather late blooming interest, the coastal ocean is a region of critical ecological and economic importance, and dynamically differs from the deep ocean in many respects. A broad coastal ocean, such as the Mid-Atlantic Bight, may be divided into a number of regions which are dynamically different from each other. For example, adjacent to the coast is a narrow band of water which is physically distinct from waters further offshore. This region is about 10 km wide and has been termed the coastal boundary layer. Scientific investigation of the coastal boundary layer is even more recent than that of the overall coastal ocean. Csanady (1977) has given a detailed review of concepts related to the coastal boundary layer. Other recent studies have been reported by Murray (1975), Bennett and Magnell (1979), Crout and Murray (1979), May (1979), Blanton (1981), Murthy and Dunbar (1981), and Pettigrew (1981).

From 1974 through 1978 a series of intensive measurements of temperature, salinity and water velocity were made within the coastal boundary layer off Long Island's southern shore. Data were both continuously recorded from moored instruments and collected by hydrographic surveys. This effort, conducted jointly by Brookhaven National Laboratory and Woods Hole Oceanographic Institution investigators, has been termed the COBOLT (COastal BOundary Layer Transect) project. This report presents analyses of the currents measured during the COBOLT project and compares these with simple theoretical concepts. Three different aspects of the motion are considered. These are: monthly means, "events" of a few days duration, and fluctuations at subdiurnal frequencies. Discussion of results will be preceded by a description of the COBOLT project and a review of relevant work by other investigators.

2. The Coastal Boundary Layer Transect Project

The COBOLT Project consisted of a number of separate field experiments, each of approximately one month duration. Most experiments involved measuring time series of temperature, salinity, and water velocity using an array of moored instruments. During all experiments, hydrographic surveys, which measured the above quantities at closely spaced flag stations, were performed on a near daily schedule. The experimental configuration and measurement techniques are only briefly described herein; refer to May (1979), Hopkins et al. (1979a, 1979b and 1982) and Pettigrew (1981) for a more detailed account.

The site of the COBOLT measurements and the arrangement of the moorings and flag stations are shown in Figures 2.1 and 2.2 respectively. The mooring array consisted of four moorings situated on a line nearly perpendicular to the coast and local isobaths. The moorings, located at 3, 6, 9 and 12 km from shore, were of Shelton spar buoy type described by Dimmler et al. (1976). Each carried four instrument packages which recorded averaged values of temperature, salinity, and orthogonal water velocity components at hourly intervals. The three outer moorings had sensor packages at depths of 4, 8 and 16 m below the mean sea surface plus one at 2.4 m above the bottom. The mooring at 3 km had packages at depths of 4, 8 and 12 m, together with one at 2.4 m above the bottom. The notation of Pettigrew (1981) will be used to refer to the instrument packages. Moorings are assigned numbers 1 to 4 progressing seaward. At each mooring, sensor package levels are given numbers 1 to 4, increasing with depth. A number pair consisting of mooring number followed by a package level number will refer to a particular sensor package. For example, the number pair (3,2) denotes the sensor package at level 2 (8 m depth) of mooring 3 (9 km from shore).

Flag stations, from which the hydrographic surveys were taken, were established along a line nearly coincident with the mooring array. They were located at 1 km intervals from 1 to 8 km from shore, and at 2 km intervals from 8 to 12 km from the coast. At each station measurements of temperature, salinity and water velocity were made from 2 m below the surface to near the bottom at intervals of 2 m.

Meteorological data were taken from a shorebased 22 meter tower near the moored array and/or from a 110 m tower at Brookhaven National Laboratory, about 35 km from the project site.

Temperature and salinity measurements of the hydrographic surveys were accurate to $\pm 0.2^{\circ}\text{C}$ and ± 0.2 ppt respectively. The uncertainties of temperature and salinity measured by the moored sensors were 0.02°C and 0.2 ppt respectively. Moored velocity measurements had uncertainties of 1 cm/s for each axis and 5° in orientation.

The mooring data used in this study were initially processed in the manner described by Pettigrew (1981). Record gaps of less than 6 hours were filled using cubic spline interpolation. To achieve a common time base, values of each time series were linearly interpolated to the nearest whole hour. Velocity components were transformed into a coordinate system with axes perpendicular and parallel to the local isobath orientation as estimated from bathymetric charts. The coordinate system used throughout this report has x positive offshore and y positive alongshore towards the northeast.

Preliminary COBOLT hydrographic surveys were conducted during June, September and October of 1974. In September of 1975, data were collected from one mooring and from the hydrographic surveys. Full deployment of four moorings together with the hydrographic surveys were carried out during May 1977, March 1978, and August 1978. Due to instrument malfunction, reliable data were collected from only three of the moorings during May 1977, and from only two moorings in March 1978. All moorings returned good data during August 1978.

Figure 2.1: The coastal region of eastern Long Island. COBOLT moorings were located at circles labeled 1 - 4.

Figure 2.2: Cross-sectional view of the experimental configuration showing the location of moored instruments and flag stations.

3. Review of Work by Other Investigators

Current meter measurements of the COBOLT experiments have been the subject of several publications. Some of these findings provide a good background for the material to be presented in the following sections and will be briefly reviewed here.

Scott and Csanady (1976) analyzed data from the September 1975 experiment which consisted of a single mooring at 11 km from shore. They found that the wind driven current, which was superimposed on a strong tidal flow, was forced primarily by the longshore component of wind stress. This non-tidal current appeared to respond promptly to the wind stress at all measurement levels, which they took as evidence of frictionally controlled flow with vigorous vertical momentum transfer. The cross-shore current response was roughly consistent with a simple "Ekman type" model with upwelling and downwelling circulation accompanying positive longshore (northeastward) and negative longshore winds respectively.

There was a marked asymmetry in the response of longshore currents to wind stress, with more windward momentum being associated with negative longshore winds than with positive longshore winds. A similar asymmetry in the response of Mid-Atlantic Bight currents had been noted by Beardsley and Butman (1974) and later, in greater detail, by Noble and Butman (1979). Despite the fact that the wind stress components averaged over the September experiment were near zero, there was an appreciable southwestward mean flow of about 5 cm/s at all levels. This is part of a body of evidence which firmly establishes the existence of a mean southwestward flow over the northeast U.S. continental shelf (Csanady, 1979). Such flow is indirect evidence of a permanent longshore pressure gradient directed downward towards the southwest.

Pettigrew (1981) considered data from all COBOLT experiments, particularly from August 1978. He found that a mean southwestward flow and the asymmetric response of longshore currents to longshore wind stress were consistently occurring features. In particular, he discovered that during positive longshore winds there was frequently a reversal in the longshore flow at about 5 km, with nearshore water moving with the wind and offshore water flowing against the wind. In contrast, when winds were in the negative longshore direction, strong southwestward flow occurred at all moorings.

Kinetic energy spectra of the longshore velocity component measured at mooring 1 during August 1978, taken from Pettigrew (1981), are shown in Figure 3.1. Pettigrew stated that these spectra are typical of those at all moorings. Note that these spectra are red, i.e. have decreasing energy with increasing frequency, and that there is a significant decrease of subtidal energy with increasing sensor depth. The dominant peak is at the semi-diurnal tidal frequency and accounts for roughly 1/4 of the total variance.

Spectra of the cross-shore velocity component at the same mooring, again from Pettigrew (1981), are shown in Figure 3.2. Note that the dominant peak is also at the semi-diurnal frequency. Energy levels are significantly less than the corresponding longshore velocity energy levels, increasingly so for decreasing frequency. This is likely a consequence of low frequency cross-shore flow being inhibited near the coast. Cross-shore energy levels do not decrease with depth. This is consistent with the multilayer character of the flow, which will be demonstrated in Sections 4 and 5.

An extensive examination of tidal currents in the COBOLT study area has been reported by May (1979). Because subtidal currents are of primary interest to this study, May's results will not be considered here.

Figure 3.1: Kinetic energy spectra of the longshore velocity component at mooring 1 during August 1978, taken from Pettigrew (1981). The curves are for sensor depths of 1, 2, and 4 m.

Figure 3.1: August 1978 alongshore kinetic energy spectra: 3 km from shore at depths of 4, 8, 12, and 18 m; from Pettigrew (1981).

Figure 3.2: August 1978 cross-shore kinetic energy spectra: 3 km from shore at depths of 4, 8, 12, and 18 m; from Pettigrew (1981).

4. Observation of Subtidal Velocities During August, 1978

Figures 4.1 and 4.2 present longshore and cross-shore velocity components of the August 1978 experiment at all current meters except (3,3) (i.e. at 9 km offshore, 16 m depth) which malfunctioned. The velocities shown have been low passed filtered using a Stallings P3 filter (McCullough, 1968) with a 26 hr cutoff. Shown in Figure 4.3 are longshore and cross-shore wind stress components (unfiltered) together with the water velocities at mooring 4. These figures illustrate several of the findings reviewed in Section 3.

The mean longshore velocity is negative (southwestward) at all moorings, despite the fact that the longshore wind stress averaged over the experiment is slightly positive. Also, the mean longshore velocity becomes increasingly negative in the seaward direction. At all current meters, fluctuations of the longshore velocity resemble and are nearly in phase with fluctuations of the longshore wind stress. Longshore velocities appear to be correlated over the length and depth of the study area.

The structure of the cross-shore velocities shown in Figures 4.2 and 4.3 appears to be a good deal more complicated than that of longshore velocities. It may be initially viewed in the framework of a two-layer flow in which the upper layer transport is "mean wind drift" (i.e. to the right of the longshore wind) and the lower layer transport is opposite to and compensates for that of the upper layer. In the coordinate system considered, a positive longshore wind stress would produce a positive (offshore) flow near the surface and a negative return flow below. Cross-shore velocity fluctuations should, according to this model, be coherent with the longshore wind stress with close to zero phase difference near the surface and with close to 180° phase difference in deep water. Evidence for this type of flow over the western U.S. continental shelf has been reported by Halpern (1976), Allen and Kundu (1978) and Winant (1980).

A close examination of the 4 m cross-shore velocities reveals that they generally resemble and are nearly in phase with the longshore wind stress. The only notable exception to this occurred August 20-21 during which there was a negative longshore wind impulse with maximum stress of about 1 dy/cm^2 accompanied by an appreciable offshore flow near the surface. This flow was

possibly accelerated by the cross-shore component of wind stress which was directed offshore with a maximum value of roughly 2 dy/cm^2 .

At all moorings the cross-shore velocity at the near-bottom current meter appears to be correlated and nearly 180° out of phase with both the 4 m velocity and the longshore wind stress. The only notable exception to this was, again, on August 20-21 during which the near-bottom cross-shore fluctuations at moorings 2, 3 and 4 were nearly in phase with the 4 m fluctuations.

Cross-shore velocity fluctuations at levels 2 and 3 are also indicative of two-layer flow, appearing to be correlated and nearly in phase with either the 4 m or near-bottom velocities. The overall pattern of cross-shore flow is, however, somewhat complicated and variable in time. The cross-shore velocity at the 16 m level of mooring 4, for example, is sometimes in phase with the 4 m velocity and sometimes in phase with the near-bottom velocity. The 16 m and 4 m velocities were in phase even on occasions when the base of the surface mixed layer was well above 16 m. These facts suggest that the apparent "mean wind drift" depth varied and did not always coincide with the mixed layer depth.

Indirect evidence of the two-layer nature of the cross-shore flow is provided by data of the hydrographic surveys. In almost every instance, lines of constant temperature, salinity and density moved upward and shoreward following a period of northeastward (upwelling favorable) wind and downward and seaward following a period of southwestward wind.

Figure 4.1: Low passed cross-shore velocities measured by moored current meters during August 1978.

Figure 4.1: Low passed longshore velocities measured by moored current meters during August 1978.

Figure 4.2: Low passed offshore velocities measured by moored current meters during August 1978.

Figure 4.3: August 1978 wind stress components at Tianna Beach, Long Island, and low passed velocity components at mooring 4.

5. Statistical Analysis

5.1 Cross-spectral analysis

Although inspection of the August 1978 time series has indicated an occasionally complicated response to the wind stress, there is nonetheless a clear relation between water velocity and the longshore wind stress. Such a relationship can be quantified by use of cross-spectral analysis.

In order that the cross-spectral calculations may provide physical insight, they will be compared with two simple analytical models which separately consider the effects of stratification and bottom friction. The models are formulated in Appendices A1 and A2. It will be instructive to examine the response predicted by each model prior to considering the cross-spectral results.

The model in Appendix A1 considers velocities of a two-layer coastal ocean which are driven by longshore wind stress. Similar models may be found in texts by Csanady (1982) and Pedlosky (1979). The upper and lower layers are assumed to be vertically homogeneous. Bottom friction and longshore variations are ignored. So that the results may be compared with spectral analysis, the wind stress and velocities are assumed harmonic in time (i.e. of the form $Ae^{i\omega t}$). Equations A1.5 through A1.8 give the longshore and cross-shore velocity components in the upper and lower layer per unit of kinematic wind stress (transfer functions). The cross-shore velocity of the upper layer is in phase with the longshore wind stress. It has zero magnitude at the coast and increases in the seaward direction on a scale of R , the internal deformation radius. The cross-shore velocity of the lower layer is opposite to that of the upper layer, and thus 180° out of phase with the longshore wind stress. The longshore velocity of both layers lags the wind stress by 90° . The longshore velocity magnitude of the upper/lower layer decrease/increase with offshore distance respectively. Near the coast the upper layer longshore velocity is much greater than that of the lower layer. Far from shore ($X \gg R$) longshore velocities of the two layers are equal.

The model of Appendix A2 gives the response of a homogeneous fluid to harmonic longshore wind stress forcing. Bottom stress is included and parameterized by a linear drag law. Because the net cross-shore transport is

considered zero, only the transfer function of the longshore velocity with the wind stress is given. Note that the phase lag of the longshore velocity with wind stress approaches 90° for large values of frequency and/or bottom depth (i.e. for $\omega H \gg C_d$). The phase lag goes to 0° for decreasing frequency and bottom depth.

Both models ignore the cross-shore wind stress component. By scaling terms in the momentum balance equations, it can be shown that, close to shore, the effect of low frequency longshore wind stress fluctuations is significantly greater than that of cross-shore fluctuations of similar magnitude. In addition, Pettigrew (1981) used COBOLT data from August 1978 to show that the cross-shore wind stress term in the cross-shore momentum balance was negligible compared with the Coriolis term. The longshore wind stress was, by contrast, comparable with all other terms of the longshore momentum balance.

These models have been purposely made simple so that features of the response can be easily identified. Because these models are so extremely idealized they are used diagnostically, i.e. to identify characteristics of the observed flow which may be attributed to either friction or stratification.

Cross-spectral results of the May 1977, March 1978, and August 1978 experiments are to be presented. The water column was highly stratified during the August and May experiments, and nearly homogeneous during the March experiment. The character of stratification during May 1977 is shown by the average σ_t field in Figure 5.1 (from Pettigrew, 1981). This shows a pycnocline roughly between 8 and 16 m (or between current meters 2 and 3) with weakly stratified water above and below. The average σ_t field of August 1978 was similar except that the pycnocline was broader with a larger density contrast between waters above and below.

Figure 4.3 shows that the wind stress during August 1978 was always small, seldom exceeding 1 dy/cm^2 . As a result, the vertical motion of isopycnals was not appreciable. Such conditions are ideal with regard to comparing measurements to the model of Appendix A1 because: (1) the model assumes small interfacial displacement relative to the layer depths; and, (2) the sensor packages were always in about the same locations with respect to

MAY 1977 AVERAGE

Figure 5.1: Twenty-two day average sigma-t field during May 1977; from Pettigrew (1981).

the isopycnals. During the May experiment there was an event with significant wind stresses which caused the water column to be homogeneous for about one day. This must be kept in mind when considering cross-spectral calculations of the May 1977 data.

Velocity measurements during March 1978 were unfortunately plagued by instrument and weather related problems which have not been completely resolved. Results of calculations involving the March data should, regrettably, be viewed with caution.

Cross-spectral estimates between velocity and wind stress components have been calculated from individual spectra which were ensemble averaged over 3 to 5 overlapping pieces. The mean and linear trend of each piece was removed. Spectral leakage was reduced by multiplying each piece by a Hanning (full cosine) window with an $(8/3)^{1/2}$ variance correction applied to the coefficients. Additional statistical stability was achieved by averaging over frequency bands.

It should be noted that frequency band averaging may or may not be appropriate depending on the character of the fluctuations. In particular, if most of the variance within a band occurs near a specific frequency, averaging over the band may lead to results which are not physically meaningful. Fortunately, the subtidal portion of COBOLT velocity spectra (e.g. Figs. 3.1 and 3.2) and wind stress spectra were always red with no statistically significant peaks or valleys (at the 80 percent significance level). Unless otherwise noted, spectral calculations presented below have been averaged over frequencies from $1/256 \text{ hr}^{-1}$ to $1/32 \text{ hr}^{-1}$. In all cases, band averaged results were qualitatively similar to those within any sub-band.

From inspection of the time series in Figure 4.3, it appears that subtidal velocity fluctuations of the August 1978 experiment were coherent with the longshore wind stress component; but, were generally not coherent with the cross-shore wind stress. This is not surprising in light of the previous discussion. These apparent relations are corroborated by cross-spectral calculations. Coherence coefficients relating subtidal August velocity components with the cross-shore wind stress differed from zero at the 95 percent significance level for only two of the 30 cross-spectra. In

contrast, coherences with the longshore wind stress differed from zero at the 95 percent level for all but four of the 30 cross-spectra.

5.1a. Relation between longshore wind stress and cross-shore velocity.

Listed in Table 5.1a are the coherence coefficients together with the amplitude, phase, and 80 percent confidence intervals of the complex transfer functions relating the longshore wind stress with cross-shore velocities measured during the August 1978 experiment. To facilitate interpretation, the coherences and transfer functions are also displayed on a transect cross-section in Figure 5.1. The transfer function phases, which are close to 0° at the 4 and 8 m levels and close to 180° at the lower two levels, indicate a two layer response to the longshore wind stress, similar to that of model A1. At each mooring, the integral over depth of the transfer function magnitudes associated with the near 0° phases is about equal to the integral of the magnitudes associated with the near 180° phases. To the degree of confidence that can be placed in a 4 point resolution, this indicates approximate 2-dimensional mass balance of the cross-shore velocity fluctuations due to longshore wind stress.

The coherences of Figure 5.1 indicate that the portion of the cross-shore energy which can be attributed to longshore wind stress was roughly 40 to 60 percent at the near-bottom current meters, but only 10 to 45 percent at those above. One reason for this difference could be that a smaller portion of the near-bottom cross-shore signal was due to mechanisms other than longshore wind, i.e. such as longshore density gradients. Also, coherences at intermediate depth current meters may be lower because these are most likely to be traversed by the interface of the "dynamic layers", causing the current meters to be sometimes in the "upper layer" and sometimes in the "lower layer". An examination of the time series has suggested that this was the reasons why the coherence was anomalously low at current meter (2,2), and why the phase was anomalously different from 180° at current meter (4,3).

As noted, phases in Figure 5.2 suggest that two "dynamic layers" made up the cross-shore response to longshore wind. Transfer function magnitudes at current meters in the "upper dynamic" layer (at 4 and 8 m) increase in the

seaward direction, similar to the response of model A1. The upper layer transfer function of this model (Eq. A1.5) has been calculated with parameters chosen so as to closely match August 1978 conditions (e.g. $\epsilon = 0.003$, $h_1 = 10$ m and $h_2 = 22$ m, which gives $R = 5$ km). Calculated transfer function magnitudes at the mooring locations are listed above the moorings in Figure 5.2. While these values are somewhat smaller than the average of the 4 and 8 m transfer function magnitudes, they nonetheless show a similar rate of increase with offshore distance.

It should be noted that a large fraction of the transfer function magnitude uncertainty is due to the determination of wind stress from wind velocity data. The formula of Amorochio and Devries (1980) has been used to calculate wind stress. This is a quadratic drag law with a drag coefficient which increases with wind speed. Within the range of wind speeds most commonly measured during August 1978 the drag coefficient is approximately 1.05×10^{-3} . Other investigators have proposed coefficients in the same speed range which are about twice as large (Neumann and Pierson, 1966). One must conclude that the uncertainty of wind stress values calculated from wind speed is very large. A close quantitative agreement between experimentally and analytically determined transfer function magnitudes should, therefore, not be expected, regardless of how elaborate the analytic model.

It is easily realized, though, that a change in the formula used to calculate wind stress would have little effect on transfer function phases or the ratio between magnitudes at different current meters. Thus, the spatial structure of transfer functions (e.g. the rate of increase with offshore distance) would be about the same regardless of the wind stress formula used.

A source of uncertainty which may affect both transfer function magnitude and phase is due to the choice of cross-shore coordinate orientation. In general, calculations involving cross-shore velocities measured close to shore should be treated suspiciously. This is because cross-shore flow near the coast is generally much smaller than longshore flow; thus, the cross-shore component of velocity may be significantly altered by a small change in the orientation chosen for the local coordinates. The orientation of the local coordinates used in this study was compared by

Pettigrew (1981) to the rotary spectral ellipse orientation of low frequency ($1/200 \text{ hr}^{-1}$) currents. The mean of the major axis orientation, which defines a dynamical shoreline, differed from the local longshore axis by only 2.5° . The orientation uncertainty of the localized velocities is thus approximately equal to the compass uncertainty of 5° . To test the sensitivity of the cross-spectral results on coordinate system orientation, calculations were done using velocities of a coordinate system which was 5° clockwise of the chosen local system. Qualitatively, the results for August 1978 were similar to those in Figure 5.2. Quantitatively, they differed by much less than the 80 percent confidence intervals.

Coherences and transfer functions relating cross-shore velocities measured during May 1977 with the longshore wind stress are shown in Table 5.1b and Figure 5.3. Similar to those of the August 1978 experiment, the phases suggest a two-layer structure of the wind driven cross-shore flow. Also similar to the August 1978 experiment, the transfer function magnitude at the two uppermost current meters increases in the seaward direction on a scale nearly equal to the estimated internal deformation radius.

If vertical isopycnal motion, as modeled by Appendix A1, were primarily responsible for the offshore increase of the August and May near-surface transfer function magnitudes, then one would expect transfer functions calculated from the March 1978 data to exhibit a much smaller increase with offshore distance. This is because there was almost no vertical stratification in March and thus $R \approx 0$ in Eqn. A1.5.

Cross-spectral values relating longshore wind stress with cross-shore velocities measured during March 1978 are shown in Table 5.1c and Figure 5.4. Unfortunately, reliable data was available from only two moorings, and coherence coefficients at the near-surface current meters are small. As for the August and May results, phases are indicative of a two-layer, mass compensating cross-shore flow. The increase of the near-surface transfer function magnitude with offshore distance does appear to be much smaller than the increase of the August and May transfer functions. Confidence intervals of the March magnitudes are, however, quite large.

Figure 5.2: Coherences together with amplitude and phase of transfer functions relating cross-shore velocity with longshore wind stress for August 1978. Mag. phase indicated wind stress leads velocity. Coherences exceeding 0.38 differ from zero at the 95 percent significance level. Transfer function is shown only where coherence is significant at the 80 percent level. Numbers above the moorings are from Eq. A1.5.

Figure 5.3: Coherences together with amplitude and phase of transfer functions relating cross-shore velocity with longshore wind stress for May 1977.

Figure 5.4: Coherences together with amplitude and phase of transfer functions relating cross-shore velocity with longshore wind stress for March 1978.

The qualitative agreement of the August and May transfer functions with the two-layer model of A1 does, by no means, prove that this model represents all important dynamical processes near the coast. Indeed, processes which should be of importance, such as bottom friction and cross-isopycnal mixing, are not included in the model. Nonetheless, it can be stated with confidence that the cross-shore flow which is statistically related to the longshore wind stress shows a two layer, mass compensating structure, and the "upper layer" fluctuations increase with offshore distance on a scale nearly equal to the estimated internal deformation radius.

5.1b. Relation of longshore wind stress with longshore velocity. Results of cross-spectral calculation relating subdiurnal longshore velocity fluctuations of August 1978 with the longshore wind stress are shown in Table 5.II and Figure 5.5. Coherence coefficients indicate that 10 to 60 percent of the longshore kinetic energy is statistically related to the longshore wind stress. Unlike transfer functions involving cross-shore velocities, those of Figure 5.5 do not resemble the model of Appendix A1. Transfer function magnitudes at near-surface current meters do not monotonically increase in the onshore direction, as the model predicts. Transfer function phases are everywhere significantly different from 90° and approach 0° with decreasing bottom depth, which is indicative of the influence of bottom friction. It is noteworthy that phases at 4 and 8 m show the largest decrease between moorings 2 and 1 where the bottom depth also decreases by the greatest amount.

5.1c. The effect of cross-shore wind stress. A storm of about three days duration with significant offshore winds occurred during the May 1977 experiment. Csanady (1980) demonstrated that the large longshore currents which accompanied this storm were likely generated by a longshore pressure gradient produced by spatial variation of the cross-shore wind stress. Consistent with this is the fact that subdiurnal longshore velocities of the May 1977 experiment were significantly coherent with both the longshore and cross-shore wind stress components. The subdiurnal offshore velocities of this experiment were generally more coherent with the longshore wind stress,

Figure 5.5: Coherences together with amplitude and phase of transfer functions relating longshore velocity with longshore wind stress for August 1978. Neg. phase indicates T_y leads v .

but were, nonetheless, often significantly coherent with the offshore wind stress.

Bendat and Piersol (1971) have shown that cross-spectral calculations between an output and a single input can give faulty results if the output is a consequence of two or more inputs which are correlated with each other. The authors presented a method by which the transfer function and coherence coefficient between the output and a single input can be calculated with the effects of the other inputs removed. Their method has been used to calculate transfer functions and coherence coefficients between COBOLT velocities and the longshore wind stress with the effect of the cross-shore wind stress removed. Because adding inputs to the system reduces the number of independent degrees of freedom, the resulting confidence intervals were much larger than those calculated with only the longshore wind stress as an input (the ordinary cross-spectrum). Because of this, it was decided to present results of the two input calculation only if they significantly differed from the corresponding results with only the longshore wind stress as input.

The one and two input results of the August and May experiments differed by much less than the 80 percent confidence intervals of the one input system shown in Tables 5.I and 5.II. All cross-spectral results presented for these experiments have, therefore, considered only the longshore wind stress.

At some current meters, coherences between the cross-shore velocity of the March experiment and the longshore wind stress were significant at the 80 percent level only when the effect of the cross-shore wind stress was removed. Transfer functions with corresponding coherence not significant at the 80 percent level were not considered suitable for presentation. Because of this, cross-spectral values of the March experiment, shown in Table 5.Ic and Figure 5.4, were calculated with the effect of the cross-shore wind stress removed.

5.2 Empirical orthogonal functions of subdiurnal velocities

Spectral analysis presented in Section 5.1 considered only velocity fluctuations which were statistically related to wind stress. Additional insight of nearshore dynamics may be gained by examining the overall velocity

fluctuations. A particularly attractive means of representing the spatial character of fluctuations is by empirical orthogonal functions (EOFs). These are a set of independent spatial functions which constitute the optimal representation of spatial and temporal data. In the frequency domain these are the eigenvectors of the cross-spectral matrix. A set of N variables is represented by N functions, each corresponding to a specified fraction of the total variance. Usually most of the variance is associated with one or two functions.

Use of this method for analyzing geophysical time series was introduced by Wallace and Dickinson (1972) and has been applied to oceanic data sets by Kundu and Allen (1976), Winant and Olson (1976), and Pettigrew (1981), among others.

For EOFs to be physically meaningful all pairs of variables involved must be highly coherent in the frequency band considered. Coherences between all pairs of subtidal longshore velocities in the same horizontal or vertical plane were significant at the 95 percent confidence level. Subtidal cross-shore velocities were significantly coherent over all horizontal planes. Coherences between cross-shore velocities of the same mooring were sometimes low, however, making EOF analysis of the vertical structure of cross-shore velocity fluctuations inappropriate.

5.2a. EOFs of cross-shore velocity. Figure 5.6 shows the highest variance EOFs of August 1978 subtidal, cross-shore velocities. Functions are given for only the 4 m and near-bottom current meters. These functions indicate that the near-surface fluctuations are nearly in phase with the longshore wind stress and nearly 180° out of phase with near-bottom fluctuations. The magnitude of the 4 m function increases with offshore distance to 9 km, and is about the same at 9 and 12 km. This suggests a coastal boundary layer of about 10 km width.

5.2b. EOFs of longshore velocity. The 4 m and near-bottom EOFs of August 1978 longshore velocities are shown in Figure 5.7. The phase of both functions relative to the longshore wind stress clearly approaches 0° with

Figure 5.6: EOFs of offshore velocities at 4 m depth and 2.5 m above the bottom for August 1978. Magnitude and phase is relative to longshore wind stress. Neg. phase if Ty leads v. At the right of each current meter block is the coherence squared of the fluctuations at the current meter with the overall function. Shown at the far right are the percent of total variance represented by the function and coherence squared with the longshore wind stress.

Figure 5.7: Same as Figure 5.6 except that it shows EOFs of Longshore velocity for August 1978.

decreasing bottom depth, indicative of frictional effects. A further indication of the influence of bottom friction is that the change in phase is largest between moorings 1 and 2 where the bottom slope is steepest. The magnitude of both functions varies little over the entire length of the transect.

Features of the EOFs of cross-shore and longshore velocities discussed above are remarkably similar to those of the transfer functions relating these velocity components to the longshore wind stress. This is further indication that the longshore wind stress was the dominant mechanism forcing low frequency (1.5-10 day period) flow in the COBOLT region during August 1978.

Figure 5.8 shows the highest variance vertical EOFs of the August 1978 longshore velocities (a separate function for each mooring). At moorings 1 and 2, the magnitude and phase at 4 m is roughly the same as at 8 m. At both moorings the magnitude at current meter 3 is significantly smaller than at current meters 1 and 2, and the phase differs by about 12° . In contrast, the function at mooring 4 has roughly the same magnitude and phase over the upper three current meters. At all moorings the magnitude at the near-bottom current meter is much smaller than at sensors above, most likely due to the influence of friction within the bottom boundary layer.

EOF magnitudes at the three upper current meters are consistent with the response predicted by the two layer model of Appendix A1. Equations (A1.7) and (A1.8) give a magnitude difference between the longshore velocities of the upper and lower layer. This difference approaches zero as offshore distance becomes large compared with R . (Remember that the estimated internal deformation radius during August 1978 is ~ 5 km.)

Vertical EOFs of longshore velocities of the May 1977 experiment are shown in Figure 5.9. The phases, in contrast to those of the August experiment, are all close to 180° . This is likely due, in part, to storm generated pressure gradients which occurred during May 1977, as discussed by Csanady (1980). The spatial structure of the magnitudes is similar to that of the August experiment. In particular, there is a significant difference between the magnitudes at current meters 2 and 3 of mooring 2. The magnitude difference between these levels is smaller at mooring 3 and negligible at

mooring 4. It is important to note that the mean pycnocline as determined by May 1977 hydrographic data (Fig. 5.1) lay between levels 2 and 3.

Vertical EOFs of March 1978 longshore velocities are shown in Figure 5.10. In contrast to the August and May EOFs, these show little variation in magnitude over the upper three current meters. Noting again that the water column was nearly homogeneous during March (and thus $R \approx 0$), this result is also consistent with the model of Appendix A1.

Comparison of the EOFs with the model of Appendix A1 should be considered with caution because of several reasons, including the following. During the August experiment the pycnocline was often broad and did not always lay between levels 3 and 4. Although the pycnocline was narrow and almost always between levels 3 and 4 during the May experiment, a significant portion of the May velocity variance occurred during a downwelling event in which the water column was homogeneous for roughly 18 hours. Nevertheless, the EOFs have unquestionably shown that the subtidal longshore velocity fluctuations have a vertical structure which is related to the stratification and changes with distance from shore.

5.3 Summary

Statistical analysis presented in this section has demonstrated that the COBOLT subdiurnal velocity fluctuations contain signatures of stratification and friction, underscoring the importance of both these factors with regard to nearshore dynamics. Frictional effects were indicated by the phase difference between the longshore velocity and longshore wind stress, which approached zero with decreasing bottom depth. The influence of stratification was indicated by the vertical EOFs of longshore velocity. Functions calculated from data of March 1978, during which the water column was homogeneous, had roughly constant magnitude over the upper three current meters. Nearshore functions calculated from data measured during stratified conditions showed a significant change in magnitude which nearly coincided with the mean pycnocline. This magnitude difference approached zero with increasing offshore distance on a scale of about 5 km.

Figure 5.8: Vertical EOFs of longshore velocity for August 1978. Magnitude and phase are relative to longshore wind stress. Neg. phase if T_y leads v . Above each mooring are the percent of the total variance represented by the function and the coherence squared with the longshore wind stress.

Cross-shore velocity fluctuations, statistically related to the longshore wind stress, showed a definite two-layer structure, with flow in the "upper"/"lower" layer being in/out of phase with the longshore stress, respectively. Magnitude of the "upper" layer flow increased with offshore distance on a scale of about 5 km. This observation, together with the vertical structure of longshore fluctuations discussed above, reveals a boundary layer character of the nearshore flow measured during stratified conditions which is similar to that of the model of Appendix A1.

Figure 5.9: Same as Figure 5.8 except that it shows vertical EOFs of Longshore velocity for May 1977.

Figure 5.10: Same as Figure 5.8 except that it shows vertical EOFs of longshore velocity for March 1978.

Table 5.1

Results of cross-spectral calculations relating cross-shore velocity with longshore wind stress (frequency band: $1/256 \text{ hr}^{-1} - 1/30 \text{ hr}^{-1}$). For each sensor coherence, transfer function magnitude and phase (- for wind stress leading) are listed from top to bottom. Confidence limits are for the 80 percent significance level. Only transfer Fns. with coh. exceeding the 80 percent level are listed.

Mooring No.	1	2	3	4
<u>5.1a. August, 1978</u>				
Level No.				
1	0.34 3.4 ± 2.7 $0^\circ \pm 35^\circ$	0.53 8.7 ± 4.1 $-4^\circ \pm 20^\circ$	0.67 15.3 ± 5.0 $12^\circ \pm 13^\circ$	0.54 12.6 ± 5.8 $23^\circ \pm 19^\circ$
2	0.30 4.2 ± 3.9 $-26^\circ \pm 42^\circ$	0.09 --- --- --- ---	0.41 5.8 ± 3.7 $49^\circ \pm 28^\circ$	0.49 8.2 ± 4.3 $51^\circ \pm 22^\circ$
3	0.55 4.9 ± 2.2 $171^\circ \pm 19^\circ$	0.56 7.1 ± 3.1 $-171^\circ \pm 18^\circ$	--- --- --- --- --- ---	0.51 6.8 ± 3.4 $139^\circ \pm 21^\circ$
4	0.62 5.4 ± 2.0 $-171^\circ \pm 15^\circ$	0.62 7.1 ± 2.7 $-172^\circ \pm 16^\circ$	0.69 8.5 ± 2.7 $167^\circ \pm 13^\circ$	0.76 6.6 ± 1.7 $-173^\circ \pm 10^\circ$

Table 5.I (continued)

Mooring No.	1	2	3	4
<u>5.Ib. May, 1977</u>				
Level No.				
1	0.65 8.2 ± 3.8 $14^\circ \pm 19^\circ$	---	---	0.62 11.6 ± 7.5 $22^\circ \pm 27^\circ$
2	0.45 5.7 ± 4.5 $50^\circ \pm 34^\circ$	0.37 5.7 ± 5.6 $23^\circ \pm 45^\circ$	0.63 14.1 ± 8.9 $24^\circ \pm 26^\circ$	
3	0.52 3.2 ± 2.1 $179^\circ \pm 27^\circ$	0.32 ---	0.57 4.5 ± 3.3 $-126^\circ \pm 31^\circ$	
4	0.06 ---	0.23 ---	0.33 ---	

5.Ic. March, 1978*

Level No.		
1	0.41 4.3 ± 5.5 $42^\circ \pm **$	0.51 6.4 ± 6.2 $-24^\circ \pm 76^\circ$
2	0.35 ---	0.66 6.8 ± 4.5 $-54^\circ \pm 41^\circ$
3	0.41 3.9 ± 4.9 $-140^\circ \pm **$	0.37 ---
4	0.69 9.9 ± 5.9 $162^\circ \pm 37^\circ$	0.74 11.4 ± 6.0 $142^\circ \pm 32^\circ$

* Effect of cross-shore wind stress statistically removed

** Undetermined

Table 5.II

Results of cross-spectral calculations relating longshore velocity with longshore wind stress for August, 1978. For each sensor coherence, transfer function magnitude and phase (- for wind stress leading) are listed from top to bottom. Confidence limits are for the 80 percent significance level. All coh. exceed the 80 percent level.

Mooring No.	1	2	3	4
Level No.				
	0.64	0.48	0.62	0.73
1	27.8 ± 9.9 -20° ± 15°	17.9 ± 9.7 -40° ± 23°	24.3 ± 9.2 -24° ± 16°	31.2 ± 8.6 -18° ± 11°
	0.55	0.35	0.48	0.58
2	20.2 ± 9.1 -19° ± 19°	12.0 ± 9.5 -40° ± 34°	14.8 ± 7.9 -49° ± 22°	19.2 ± 7.9 -44° ± 17°
	0.54	0.49	---	0.67
3	15.8 ± 7.3 -5° ± 19°	9.7 ± 5.1 0° ± 22°	--- --- --- ---	21.3 ± 7.1 -45° ± 14°
	0.51	0.54	0.67	0.75
4	8.5 ± 4.3 -10° ± 21°	8.2 ± 3.7 -19° ± 19°	9.6 ± 3.2 -31° ± 14°	9.3 ± 2.4 -31° ± 11°

6. Monthly Mean Flow

Time averaged velocity and wind stress components of COBOLT experiments are listed in Table 6.I.

Mean longshore velocities of all experiments are directed to the southwest with generally little shear over the upper three current meters. Clearly, these mean velocities cannot be attributed to wind forcing because mean wind stresses of these experiments are either negligible or would tend to drive a northeastward flow. Csanady (1976) demonstrated that a mean southwestward flow, such as that observed, would result from a persistent longshore elevation gradient directed downward towards the southwest.

Mean longshore velocities at near-bottom current meters are much smaller than those above presumably due to bottom friction.

Despite the negligible mean wind stresses of the August 1978 and September 1975 experiments, near-surface mean cross-shore velocities of these experiments were of significant magnitude and directed offshore. Those of the August experiment are shown in Figure 6.1. These velocities cannot be attributed to instrument error which is roughly 1 cm/s. Neither can they be due to the 5° uncertainty in determining cross-shore orientation, because the local coordinates would have to be rotated by as much as 45° to make mean offshore velocities at some current meters zero. One must conclude, therefore, that there was significant offshore flow within the upper 15 m during the August and September experiments.

From measurements off Long Island in November of 1974, Han and Mayer (1981) also found appreciable near-surface, offshore flow which could not be attributed to wind forcing. Taken together, the COBOLT measurements and those of Han and Mayer show a tendency for non-wind driven, offshore flow to occur in the near-surface waters off Long Island during the late summer and fall.

It is of interest to investigate possible causes of this flow. Csanady (1976) has identified individual components which generate mean circulation in shallow water. Those which may drive a cross-shore flow in the absence of mean wind stress are: a longshore elevation gradient, a cross-shore density gradient caused by a surplus of warm and fresh water near the coast, and a longshore density gradient.

Figure 6.1: Mean cross-shore velocities of the August 1978 experiment. Mean wind stress components are shown above the moorings.

A longshore elevation gradient, such as would drive the observed mean longshore velocities, would cause a near-surface onshore flow of roughly 1 cm/s. It, thus, cannot be used to explain the mean offshore flow.

Csanady (1976) defined a density scale, $L = \rho [\partial\rho/\partial x]^{-1}$, which is the governing parameter in determining offshore flow due to cross-shore density gradient, $\partial\rho/\partial x$. Hydrographic sections within the Mid-Atlantic Bight indicate that this parameter has a maximum value in the vicinity of the Hudson which is about 10^{10} cm. This value, according to Csanady's (1976) formulation, gives an offshore flow of ~ 1 cm/s. Density sections of the August and September COBOLT experiments indicate L values much smaller than 10^{10} cm. Thus, the mean offshore flow near Long Island which results from cross-shore density gradients is probably negligible.

This leaves longshore density gradients as the only remaining candidate for explaining the observed mean offshore flow. Available hydrographic data collected during the late summer and fall consistently exhibit longshore density gradients near Long Island which would tend to generate offshore flow. An example of such a gradient is contained in Figures 6.2 and 6.3 which show a longshore hydrographic section taken close to Long Island during August 19-22, 1978. The isopycnal slope over the entire transect is downward towards the southwest which would tend to produce offshore drift.

While it is difficult to compute absolute density driven flow in shallow water, relative geostrophic velocities can be estimated by the thermal wind equation:

$$u(z_2) - u(z_1) = \frac{-g}{\rho f} \int_{z_1}^{z_2} \frac{\partial \rho}{\partial y} dz \quad (6.1)$$

where:

- u = offshore velocity
- y, z = longshore/vertical coordinate
- g = gravity
- f = Coriolis parameter
- ρ = density

Figure 6.2: Hydrographic stations of the Russian research vessel BELOGORSK, taken August 18-22, 1978.

Figure 6.3: Temperature, salinity and density section parallel to Long Island. Station locations shown in Figure 6.2.

Figure 6.1 indicates that the 4 m mean cross-shore velocity of the August experiment was 4-8 cm/s offshore relative to the 16 m mean velocity which was nearly zero. The 4 m offshore velocity relative to 16 m, computed by Eq. 6.1 with the density gradient between stations 64 and 65 of Figure 6.2 (which straddle the COBOLT transect), is ~ 3 cm/s. This value, while lower than the observed mean difference between 4 and 16 m, is nonetheless of the same order of magnitude. It will be shown in Section 7 that the longshore density gradient near the COBOLT site was at times during August 1978 much greater than that shown in Figure 6.3. Thus, longshore density gradients may well have been responsible for nearly all the mean offshore flow of the August 1978 experiment. Such gradients, which are frequently apparent in late summer and fall hydrographic data near Long Island, may likewise have generated the mean offshore flow of the September 1975 experiment and that measured by Han and Mayer (1981). Hydrographic data taken during March and May of 1978 (from Pantanjo *et al.*, 1982) show no appreciable longshore density gradients near Long Island. Significant density driven offshore flow during these months would, thus, seem unlikely. This supposition is corroborated by the March 1978 and May 1977 averaged velocities given in Table 6.1. Mean wind stress components of the March experiment were near zero and the mean offshore velocities differed from zero by less than the instrument error. Near-surface mean offshore velocities of May 1977 can be attributed to the significant mean stress components, both of which would tend to generate near-surface offshore flow.

To summarize, hydrographic and current meter data taken near Long Island indicate a tendency for density driven offshore flow to occur during late summer and fall, but not during spring. It is of interest to speculate as to the cause of this apparent seasonal flow.

An obvious first guess is that the longshore density gradient responsible for the flow arises from a difference in the total fresh water discharge at either end of Long Island. The Connecticut and Hudson Rivers are, by far, the largest sources of fresh water to the Mid-Atlantic Bight (Ketchum and Keen, 1955). The Hudson's mouth is at the western end of Long Island, while the Connecticut empties into Long Island Sound near the Island's

eastern tip (Fig. 6.2). It is readily surmised that an excess of Hudson relative to Connecticut River discharge would tend to produce a longshore density gradient as shown in Figure 6.3. Figure 6.4 compares the monthly Hudson and Connecticut River discharges for 1978. (The data were kindly supplied by the Geological Survey, U.S. Department of Interior Offices at Albany, New York and Hartford, Connecticut.) Throughout the entire year, the monthly discharge of the two rivers nearly equal each other.

The effect of the Connecticut's discharge on the density structure over the Mid-Atlantic Bight is, however, determined by the water exiting Long Island Sound. Studies of Long Island Sound by Riley (1952, 1956) have shown that the Connecticut's discharge initially flows westward into the Sound and is subject to vigorous vertical mixing before eventual discharge into shelf water. Due to extremely low river discharge, the flow out of Long Island Sound during the summer and fall is vertically mixed and only slightly less saline than shelf water (Riley, 1956). This is consistent with the salinity section of Figure 6.3. During spring, when the fresh water discharge is maximum, the flow exiting Long Island Sound is vertically stratified and much fresher than adjacent shelf water. This seasonal variation in the water leaving Long Island Sound may well contribute to the observed seasonal variation of the longshore density gradient near Long Island, and, in turn, to the apparent seasonal offshore flow.

Another possible reason for the observed summer and fall time excess of heavy water near the eastern end of Long Island is an inflow of dense shelf water into the region. Bumpus (1973) compiled ten years of surface drifter data over the continental shelf into monthly composite flow diagrams. His diagrams for July, August and September show an onshore flow slightly to the east of Long Island (and a bit west of Nantucket Shoals). They also show sporadic offshore flow near Long Island, and longshore flow in the vicinity of the Hudson. The drifter data suffer a great many uncertainties, as Bumpus himself ruefully admits. Nonetheless, the data offer additional corroboration of the existence of a near-surface offshore flow near Long Island during the late summer and fall. Furthermore they indicate an influx of shelf water to the east of Long Island which may be partly responsible for the offshore flow.

Figure 6.4: Comparison of Hudson and Connecticut River monthly mean discharge rates for 1978.

Table 6.I
mean values ($\bar{u}, \bar{v}$) cm/s

Mooring No.	1	2	3	4
<hr/>				
August, 1978	$(\bar{T}_x, \bar{T}_y) = (-0.07, 0.06) \text{ (cm/s)}^2$			
Level No.				
1	(4.7, -8.7)	(2.6, -12.9)	(5.0, -13.1)	(8.1, -14.8)
2	(-3.0, -8.3)	(-0.2, -12.4)	(1.8, -14.5)	(2.5, -16.9)
3	(0.1, -8.4)	(-1.0, -14.3)	---	(0.7, -16.2)
4	(0.4, -2.9)	(1.7, -4.5)	(-0.4, -3.4)	(1.9, -2.0)
<hr/>				
March, 1978	$(\bar{T}_x, \bar{T}_y) = (-0.03, 0.10) \text{ (cm/s)}^2$			
Level No.				
1	(-0.4, -8.8)		(1.2, -4.2)	
2	(0.6, -7.0)		(-0.2, -4.7)	
3	(1.9, -9.5)		---	
4	(0.7, -2.9)		(-2.3, -1.7)	
<hr/>				
May, 1977	$(\bar{T}_x, \bar{T}_y) = (0.55, 0.39) \text{ (cm/s)}^2$			
Level No.				
1		(4.1, -5.0)	---	(3.9, -12.4)
2		(1.9, -6.8)	(-0.4, -10.3)	(3.9, -10.8)
3		(-1.6, -3.3)	(-1.5, -6.5)	(-2.2, -7.7)
4		(-2.2, -0.8)	(-1.5, -2.5)	(0.2, -2.6)
<hr/>				
September, 1975	$(\bar{T}_x, \bar{T}_y) = (0.07, -0.07) \text{ (cm/s)}^2$			
Level No.				
1				(3.2, -5.0)
2				---
3				(-1.2, -5.9)
4				(-0.3, -3.2)
<hr/>				

7. Event Analysis

Sections 5 and 6 have considered statistical properties of flow measured over a number of episodes. Additional insight may be gained by examining flow characteristics during particular episodes. In this section analysis of measurements taken during three separate events of August 1978 will be presented. These events were defined as periods during which the wind forcing was of roughly constant magnitude and direction. While this presentation considers only a small subset of all such events contained in COBOLT measurements, it, nonetheless, brings to light characteristics which are typical of many events.

7.1 Event 1: August 3-10, 1978

The longshore wind stress during this event was relatively small and nearly always in the upwelling favorable direction (Fig. 4.3). Offshore velocities and wind stress components averaged over this event are shown in Figure 7.1. Note that all mean velocities at mooring 4 are directed offshore. If the total transport at mooring 4 was offshore, continuity of mass would require an onshore, balancing flow somewhere else along the coast. Mean velocities at moorings 1 and 2, however, suggest possible unmeasured onshore flow between levels 3 and 4 at mooring 4. A further indication of such a flow can be surmised from Figure 7.2 which is a plot of daily mean density against daily mean offshore velocity at each sensor during the period of August 3-8 (excluding August 5, during which the longshore wind was briefly downwelling favorable). There is a clear separation between onshore and offshore flow near $22 \sigma_t$, with offshore velocities at lower density and predominantly onshore velocities at greater density.

The largest onshore velocities occur between 22.3 and $23.5 \sigma_t$. The one offshore velocity within this density range was measured at mooring 1. For many events, the pattern of offshore flow vs. density at mooring 1 differed from that of the other moorings. This is probably related to the shallow depth at mooring 1 and its proximity to the shore. All values with $\sigma_t > 24$ were measured at near-bottom current meters. The relatively small magnitude and occasional positive sign of these velocities can be attributed to frictional effects (i.e. Ekman veering).

Figure 7.1: Mean cross-shore velocities and wind stress components of Event 1.

The density range associated with onshore velocities (22 to $24\sigma_t$) roughly defines the pycnocline. Hydrographic data show the pycnocline at mooring 4 to lie between levels 3 and 4 throughout most of event 1. As can be seen in Figure 7.2, the onshore flow within the pycnocline was seldom measured at mooring 4. The velocity values between 22 and $24\sigma_t$ indicate that the unmeasured mean onshore velocity at mooring 4 may easily have been close to 7 cm/s, which is the value needed to make the Event 1 vertically integrated flow at the mooring equal zero (as determined by trapezoidal integration).

The possibility of significant cross-shore flow not being resolved by such a densely packed array of current meters is disconcerting; particularly so when one considers that vertical integration of the velocities in Figure 7.1 would give large and unreal offshore transport at 9 and 12 km. The "discovery" that cross-shore flow of a certain direction is often confined within a density band was fortunate in that it has provided a mechanism for finding and estimating unresolved flow at a particular mooring.

The "upper" layer during Event 1 may be considered as the band in which cross-shore flow was directed offshore. Temporally and vertically averaged wind drift velocity within this layer may be estimated by:

$$U_{WD} = \frac{\overline{T_y}}{f \overline{h_1}} \quad (7.1)$$

where: $\overline{T_y}$ is the mean longshore wind stress and $\overline{h_1}$ is the mean "upper" layer depth. Using $\overline{h_1} = 16$ m, which was a typical "upper" layer depth at 12 km, gives $\overline{U_{WD}}$ (Event 1) ≈ 2 cm/s. This value is roughly 4 cm/s less than the vertically averaged mean velocity measured within the "upper" layer (upper three current meters) at mooring 4. The large magnitude of this difference is fairly certain indication that non-wind driven cross-shore flow occurred during Event 1.

To determine if the non-wind driven flow may be attributed to longshore density gradients, hydrographic data taken in the COBOLT vicinity during August 2 and 3 as part of the MESA (Marine EcoSystems Analysis) project have

Figure 7.2: Daily average of density against daily average of offshore velocity for August 3, 4, 6, 7, and 8, 1978. For clarity values from sensor package (1,4) are not shown.

been examined. Figure 7.3 shows locations of MESA stations which were offshore of the COBOLT transect. In Figure 7.4 cross-shore geostrophic velocity between three separate pairs of stations is shown. Each pair was oriented approximately along isobaths. Although there is some variation in these profiles, they all show 2-4 cm/s offshore velocity at the surface relative to 16 m. Comparing these with the mean velocities in Figure 7.1 suggests that density driven flow could have comprised about half of the Event 1 offshore transport in the upper 16 m, and roughly accounted for the difference between the total offshore transport and that driven by the wind.

7.2 Event 2: August 11 and 12, 1978

Longshore wind stress during Event 2 was slight and downwelling favorable. Offshore velocities and wind stresses averaged over this event are shown in Figure 7.5. Vertically integrated cross-shore transport at all moorings is approximately zero. The offshore flow vs. density plot of this event, shown in Figure 7.6, is similar to that of Event 1 in that the onshore velocities fall within a definite density range and the pattern at mooring 1 is different from that of the others.

With no horizontal divergence, vertical mixing, or longshore density variation, the cross-shore flow through a shore-parallel vertical plane should be nearly equal to that inferred from movement of isopycnals in a perpendicular, or cross-shore, plane. Considering events with little net longshore transport, a comparison of the observed mean cross-shore velocities of such events with cross-shore velocities inferred from isopycnal motion gives an indication of the importance of vertical mixing and/or horizontal divergence. Throughout lengthy events, such as Event 1, isopycnal positions are likely to change due to longshore advection alone because of the significant longshore density variation in the region, as shown in Figure 6.3.

The longshore transport throughout Event 2 was small (Fig. 4.1) and probably did not significantly alter the density structure of the transect plane. Mean velocities at 12 km, inferred from isopycnal motion between the August 11 and August 13 hydrographic surveys are: 3 cm/s onshore in the "upper" layer, and 3 cm/s offshore in the "lower" layer. These are about

equal to the 12 km mean velocities of Figure 7.5, suggesting little cross-isopycnal mixing during Event 2.

This certainly does not imply that vertical mixing was insignificant throughout the August experiment. Pycnocline erosion during downwelling circulation is less likely than at other times because downwelling advects the pycnocline away from generally high turbulence regions near the shore and sea surface. For most upwelling events there was a significant difference between the measured cross-shore flow and that inferred from isopycnal motion. Significant longshore transport during these events, however, precluded using the difference to indicate pycnocline erosion.

7.3 Event 3: August 13 - 19, 1978

Longshore winds during Event 3 were light and upwelling favorable. Mean cross-shore velocities and wind stress components of this event are shown in Figure 7.7. Mean velocities at moorings 2-4 are offshore at all but one current meter. The daily average σ_t versus cross-shore velocity plot of this event shows no indication of appreciable, unresolved onshore flow. It must, therefore, be concluded that, seaward of 6 km, there was significant net offshore transport during Event 3. For mass continuity, this transport must be balanced by onshore flow somewhere else along the coast. Very little of the offshore flow of Event 3 can be attributed to wind forcing (~ 1 cm/s in the "upper" layer according to Eq. 7.1). As for Event 1, longshore density gradients may well be responsible for the bulk of the cross-shore flow.

It is interesting that mean velocities of the August 1978 experiment, shown in Figure 6.1, vertically integrate to give significant offshore transport at 12 km. Analysis of Event 1 has indicated significant unresolved onshore flow between levels 3 and 4 at 12 km. However, little unresolved flow can be surmised from the data of Event 3 and of other events during August 1978. It must be concluded that, although there was some unresolved onshore flow during August 1978, the mean cross-shore transport at 12 km was significant and offshore.

Figure 7.3: Hydrographic stations of the R/V KELEZ, taken August 2-3, 1978.

Figure 7.4: Computed geostrophic velocity profiles (solid lines), relative to the bottom, between pairs of stations shown in Figure 7.3.

Figure 7.5: Mean cross-shore velocities and wind stress components of Event 2.

Figure 7.6: Daily average density against daily average offshore velocity for August 11 and 12.

Figure 7.7: Mean cross-shore velocities and wind stress components of Event 3.

8. Summary

One interesting finding of this study concerns mass balance in the cross-shore plane. Cross-spectral analysis suggested approximate 2-dimensional mass balance of 1.5 - 10 day period fluctuations of the cross-shore flow due to the longshore wind stress. In contrast, mean cross-shore transport of the August and September experiments was significant and directed offshore. This mean flow was not generated by the wind, but was apparently due to a seasonal longshore density gradient. Surface drift data of Bumpus (1973) suggest that this observed transport off Long Island may be balanced by an influx of shelf water somewhat to the west of Nantucket Shoals. The cross-shore flow of single events was sometimes in mass balance and sometimes not. A large portion of the cross-shore flow which occurred during events with mass imbalance was evidently caused by longshore density gradients.

Examination of subtidal fluctuations has revealed a coastal boundary layer with a width about equal to that of the transect. Some of the boundary layer characteristics appear to be largely controlled by stratification and qualitatively agree with a simple two-layer coastal flow model. Because of the shoreward decrease in bottom depth, effects of bottom friction also tend to form a coastal boundary layer, as indicated by the phase of longshore currents with respect to longshore wind stress.

During many events cross-shore flow of a particular direction was confined within a density band. This provided a mechanism for detecting unresolved flow at a particular mooring which, surprisingly so, was sometimes very large.

Acknowledgements

The COBOLT hydrographic surveys were supervised by Mr. B. Pade of Woods Hole Oceanographic Institution. Dr. Thomas Hopkins and his team at Brookhaven National Laboratory were responsible for measurements from moored instruments. Dr. Paul May initially processed much of the COBOLT data. Special thanks go to Dr. Neal Pettigrew whose Ph.D. dissertation provided the background and starting point for this study. Throughout my investigation Dr.

Pettigrew was generous with assistance and advice. Drs. David Chapman and Kenneth Brink lent their expertise on theoretical aspects of coastal circulation. Finally, I am indebted to Dr. Gabriel Csanady for his patient advice and support.

This work was supported by the Department of Energy under contract number DE-AC02-79EV10005.

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APPENDIX A1

Two-Layer Coastal Jet Model

This model considers nearshore flow of a two-layer water mass which is driven by longshore wind stress. The idealized system is diagrammed in Figure A1.1. The ocean bottom is assumed to be of constant depth. The density and static equilibrium depth of the layers are denoted by ρ_1 , h_1 and ρ_2 , h_2 , where subscript 1/2 denotes the upper/lower layer. Displacement of the free and internal surfaces from their static equilibrium position is symbolized by δ and η respectively.

Only small displacements of the free surface and interface are considered (i.e. $h_1, h_2 \gg \delta, \eta$). The system is assumed to be linear, i.e. Rossby number $\ll 1$. Bottom and internal friction, and variations in the longshore direction are ignored. Interfacial displacement is considered to be much greater than that of the free surface ($\delta \ll \eta$).

With the above assumptions, equations of motion and continuity of the upper layer are:

$$\begin{aligned}\frac{\partial v_1}{\partial t} + f u_1 &= T_y/h_1 \\ \frac{\partial u_1}{\partial t} - f v_1 &= -g \frac{\partial \delta}{\partial x} \\ h_1 \frac{\partial u_1}{\partial x} - \frac{\partial \eta}{\partial t} &= 0\end{aligned}\tag{A1.1}$$

where:

x is the offshore coordinate.

t is time.

v_1 is depth averaged longshore velocity.

u_1 is depth averaged offshore velocity.

f is the Coriolis parameter.

T_y is the longshore wind stress.

g is gravity

Figure A1.1: Idealized two-layer model cross-section.

Similar equations for the lower layer are:

$$\frac{\partial v_2}{\partial t} + f u_2 = 0$$

$$\frac{\partial u_2}{\partial t} - f v_2 = -g \left(\frac{\partial \delta}{\partial x} + \epsilon \frac{\partial \eta}{\partial x} \right) \quad (A1.2)$$

$$h_2 \frac{\partial u_2}{\partial x} + \frac{\partial \eta}{\partial t} = 0$$

where $\epsilon = \frac{\rho_2 - \rho_1}{\rho_2}$

It can be assumed, without loss of generality, that the wind stress time series is a superposition of harmonic contributions, each with a characteristic amplitude and frequency, e.g.

$$T_y(t) = \sum_{k=1}^N T_y(\omega_k) e^{i\omega_k t}$$

The response to a particular contribution can be determined by substituting $T_y(\omega) e^{i\omega t}$ (subscript dropped for convenience) for T_y in (A1.1) and assuming that all other quantities are harmonic at frequency ω . Because (A1.1) and (A1.2) are linear, the total velocity at either level is simply the sum of all harmonic contributions. The model results are thus directly comparable with spectral analysis. With these substitutions (A1.1) and (A1.2) become:

$$\begin{aligned} i\omega v_1 + f u_1 &= T_y/h_1 \\ i\omega u_1 - f v_1 &= -g \frac{\partial \delta}{\partial x} \\ h_1 \frac{\partial u_1}{\partial x} - i\omega \eta &= 0 \end{aligned} \quad (A1.3)$$

and

$$i \omega v_2 + f u_2 = 0$$

$$i \omega u_2 - f v_2 = -g \left(\frac{\partial \delta}{\partial x} + \epsilon \frac{\partial \eta}{\partial x} \right) \quad (A1.4)$$

$$h_2 \frac{\partial u_2}{\partial x} + i \omega \eta = 0$$

(A1.3) and (A1.4) constitute a complete set of coupled differential equations which can be solved by applying boundary conditions of a finite solution at large X and zero cross-shore flow at the coast ($u_1, u_2 = 0$ at $X = 0$). The technique of solution is standard and need not be discussed here. At frequencies small compared with the inertial frequency ($f \gg \omega$) the velocities per unit wind stress are:

$$\frac{u_1}{\tau_y} = \frac{h_2 (1 - e^{-X/R})}{f h_1 (h_1 + h_2)} \quad (A1.5)$$

$$\frac{u_2}{\tau_y} = \frac{- (1 - e^{-X/R})}{f (h_1 + h_2)} \quad (A1.6)$$

$$\frac{v_1}{\tau_y} = \frac{e^{-i \pi/2}}{\omega (h_1 + h_2)} \left(1 + \frac{h_2}{h_1} e^{-X/R} \right) \quad (A1.7)$$

$$\frac{v_2}{\tau_y} = \frac{e^{-i \pi/2}}{\omega (h_1 + h_2)} (1 - e^{-X/R}) \quad (A1.8)$$

where R is the internal radius of deformation given by:

$$R = \frac{(\epsilon g h')^{1/2}}{f} \quad h' = h_1 h_2 / (h_1 + h_2)$$

APPENDIX A2

Longshore Flow with Linear Bottom Drag

This model considers longshore flow of a homogeneous water mass driven by the longshore wind. Assume that sufficiently close to a straight coastline, longshore pressure gradient and the net cross-shore transport are both zero. The depth integrated longshore momentum balance is:

$$H \frac{\partial v}{\partial t} = T_y - B_y \quad (A2.1)$$

where:

- v is the depth averaged longshore velocity
- H is the bottom depth
- T_y is the longshore wind stress
- B_y is the bottom stress

Bottom stress is to be parameterized by a drag law which is linear with longshore velocity.

$$B_y = C_d v$$

where C_d is a drag coefficient referred to the depth averaged velocity. A linear rather than a quadratic drag law has been chosen in order to make the results simple. As in Appendix A1, a solution will be sought for one harmonic component of the wind stress. The momentum balance is:

$$i \omega H v = T_y - C_d v \quad (A2.2)$$

Solving for v/Ty :

$$G = v/Ty = \frac{C_d - i \omega H}{C_d^2 + \omega^2 H^2} \quad (A2.3)$$

Note that the phase of G approaches zero for small H and/or ω , and approaches -90° for $\omega H \gg C_d$.

REPORT DOCUMENTATION PAGE	1. REPORT NO. WHOI-84-14	2.	3. Recipient's Accession No.
4. Title and Subtitle Analysis of Flow Within the Coastal Boundary Layer Off Long Island, New York.			5. Report Date April 1984
7. Author(s) J. H. Churchill			6.
9. Performing Organization Name and Address Woods Hole Oceanographic Institution Woods Hole, Massachusetts 02543			8. Performing Organization Rept. No. WHOI-84-14
12. Sponsoring Organization Name and Address Department of Energy			10. Project/Task/Work Unit No.
			11. Contract(C) or Grant(G) No. (C) DE-AC02-79EV10005 (G)
			13. Type of Report & Period Covered Technical
			14.
15. Supplementary Notes This report should be cited as: Woods Hole Oceanog. Inst. Tech. Rept. WHOI-84-14.			
16. Abstract (Limit: 200 words) From 1974 through 1978 a series of intensive measurements were made in the coastal waters within 12 km of Long Island. The data were derived from two sources: a mooring array from which time series of temperature, salinity and water velocity were measured at four depths at each of four offshore distances; and high resolution, daily hydrographic surveys. Analysis of subtidal cross-shore velocity fluctuations has indicated a two-layer response to wind forcing, with near-surface flow to the right of the longshore wind and opposing flow below. The magnitude of these fluctuations increased in the seaward direction on a scale nearly equal to the internal deformation radius. The phase between longshore velocity fluctuations and longshore wind stress approached zero with decreasing bottom depth, probably the result of bottom stress. The vertical structure of longshore fluctuations during stratified conditions markedly differed from that during unstratified conditions, and resembled the structure derived from a simple two-layer coastal flow model. Significant mean offshore flow was measured during experiments in August and September, despite negligible mean wind stress during the same periods. This flow was most likely due to persistent longshore density gradients, as are consistently inferred from hydrographic data taken in the vicinity.			
17. Document Analysis a. Descriptors 1. nearshore currents 2. coastal boundary layer 3. coastal dynamics b. Identifiers/Open-Ended Terms c. COSATI Field/Group			
18. Availability Statement: Approved for public release; distribution unlimited.		19. Security Class (This Report) UNCLASSIFIED	21. No. of Pages 73
		20. Security Class (This Page)	22. Price

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