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The circulation near the head of Chesapeake Bay

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ABSTRACT

Three-month long current records are examined for a position near the head of the Chesapeake Bay. For time scales longer than 5 days the flow was determined by the strength of the Susquehanna discharge. A major event was Tropical Storm Eloise which occurred near the middle of the observation period and whose influence dominated the current spectra. Wind forcing was important at time scales around 3 days and resulted in a barotropic response; both local and far-field effects had to be considered when interpreting the wind-driven response. The N-S wind caused a local response but with a significant phase lag between the wind and current; the response to the E-W wind was nonlocal and was the result of an Ekman flux in the Mid-Bay; this caused the longitudinal net flow at the mooring position to appear to flow against the wind.

The horizontal salinity gradient at the downstream limit of the freshwater zone was an order of magnitude larger than the gradient along the main portion of the Bay. During the first part of the field study this front lay upstream of the mooring position and an estuarine circulation was measured on the seaward side of it. Following Tropical Storm Eloise the front was displaced further downstream and the recorded net flow was seaward at all depths. The location of the front determines the character of the circulation at the head of the Bay and plays an important role in the maintenance of the turbidity maximum.

1. Introduction

Although Chesapeake Bay has been studied extensively during the past 25 years, there are still fundamental questions concerning the characteristics of its internal circulation and salinity distribution which remain unanswered. In particular, these problems involve the variability, the time scales and the spatial coherence of the salinity and velocity fields. Much of the work which has been conducted in the past has involved either discrete measurement or else relatively short periods (1-10 days) of continuous measurement (e.g. Klepper, 1972). The duration of such records was generally of a few tidal cycles, and the total record was usually averaged to a single number which was taken to represent the net (nontidal) mean.

In addition, while considering only steady state processes, previous investigations into the estuarine circulation within Chesapeake Bay have mainly concentrated on

the processes taking place in the well developed region of an estuary near its mouth. However, for many problems: the spawning of fish, the introduction of urban and industrial waste, the regulation of shipping channels, the influx of riverborne sediments, and the cooling of power plants, knowledge of the circulation patterns near the head of an estuary becomes important.

Recently it has been shown that meteorological forcing is important in the Chesapeake estuaries and that time scales of 2-20 days may be dominant (Elliott, 1978; Wang and Elliott, 1978; Wang, 1978) and it is appropriate to give a summary of the main conclusions. For time scales longer than 10 days a significant proportion of the fluctuations within the Bay are generated at the Bay mouth by the action of the alongshore coastal winds. The winds blowing parallel to the coastline cause an Ekman flux in the shelf waters and corresponding fluctuations in sea level at the Bay mouth. These fluctuations then propagate up the Bay. At time scales between 4-10 days coastal effects are also important, as are the lateral (E-W) winds which cause an Ekman transport within the waters of the Mid-Bay; this results in a decrease (increase) of sea level within the Upper Bay when the wind blows toward the east (west). For time scales shorter than 4 days the dominant fluctuations were found at 2-2.5 days and were interpreted in terms of a longtitudinal seiche motion.

Comparable results have been obtained in other estuarine systems (e.g. Weisberg, 1976; Smith, 1977), showing that we need to obtain a better understanding of the relation between the meteorological forcing and the nontidal circulation. To resolve such questions it is necessary to obtain continuous current records which are much longer than the 5-10 days of previous studies and which are sufficiently long that they can be used to describe the time varying nature of the residual flow. Toward this goal, this paper presents the results of a three-month long investigation conducted near the head of Chesapeake Bay and discusses the effects of wind stress and river flow on the observed currents.

2. The data

A mooring containing two current meters was maintained in the Upper Chesapeake Bay near Howell Point (Fig. 1). This region is located on the Eastern Shore about 270 km from the Bay mouth and 25 km downstream from the mouth of the Susquehanna at Havre de Grace. Water depth in this section of the Bay is generally less than 5 m, although there is a shipping channel toward the Eastern Shore which has depths of 7.5-12.0 m. The mooring was located in water 6.5 m deep near the edge of the channel, approximately 1300 m to the SW of Howell Point. The two current meters, moored at depths of 2.1 m (7 feet) and 4.6 m (15 feet) (the data series will be referred to as U_7 and U_{15}), recorded during a three-month period from August 4 to October 30, 1975.

Wind data were obtained from Baltimore-Washington International Airport which is approximately 55 km to the SW of the mooring position, and from the Patuxent
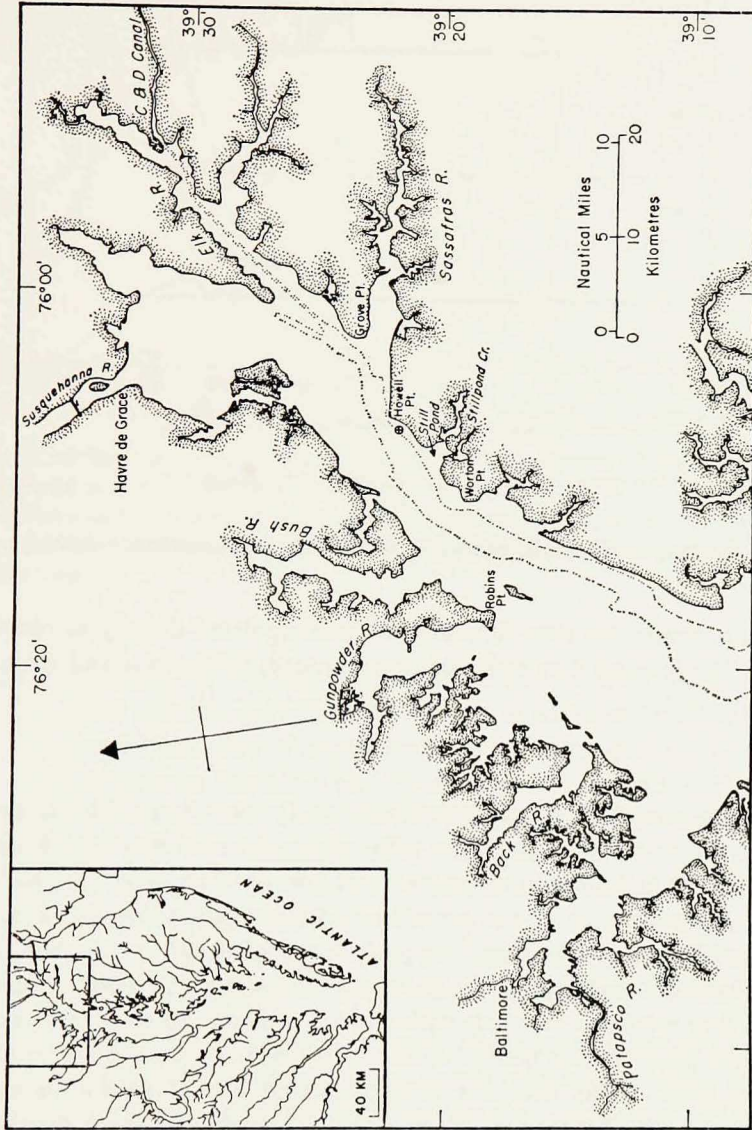
Figure 1. The Upper Chesapeake Bay showing the mooring location at 39-21-54N, 76-07-18W. (The 18 feet isobath is marked by the dotted line.)

Naval Station which lies 125 km to SSW of the mooring. The hourly values of wind data were converted to wind stress by using a quadratic law. Flow data for the Susquehanna River were obtained as mean daily discharges at the Conowingo Dam which lies 5 km upstream of Havre de Grace. In addition, sea level data were obtained from Havre de Grace and Baltimore Harbor (Fig. 1). The data were lowpass filtered to remove the tidal and other high frequency components and then resam-
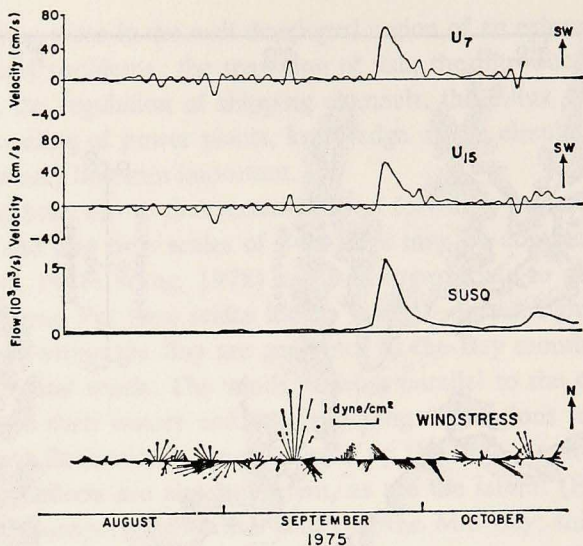
Figure 2. The lowpass time series of current, wind stress and Susquehanna discharge.

pled at 6-hour intervals; the river flow data were interpolated directly to obtain 6-hourly values. The lowpass time series of current speed, wind stress and river flow are shown in Figure 2.

3. Results

During the first part of the observations (prior to Tropical Storm Eloise) the net circulation showed an estuarine character with a mean seaward flow of 1.6 cm s^{-1} near the surface (U_7) and a return mean landward flow of -2.7 cm s^{-1} at the lower level (U_{15}). The second part of the experiment was dominated by the effects of Tropical Storm Eloise when the river flow peaked at 16,530 m^3 s^{-1} (on September 27) and caused net seaward flows of the order of 60 cm s^{-1} at the mooring position. The mean net flow remained seawards with a speed of about 10 cm s^{-1} during the remainder of the measurement period. Although the winds were not severe during Eloise, several other events occurred when the wind caused significant current fluctuations; the largest of these took place around September 12 when there was a strong northward wind stress. Since the peak flow associated with Eloise dominated the current records, only the first part of the data series (August 4 to September 23) was used to analyze the effects of wind forcing; the total records were used for the other calculations.

The spatial structure of the wind. This section examines whether the wind data collected at the distances of 55 km and 125 km from the mooring position can be taken as being representative of the lowpass winds over the entire area. Figure 3a
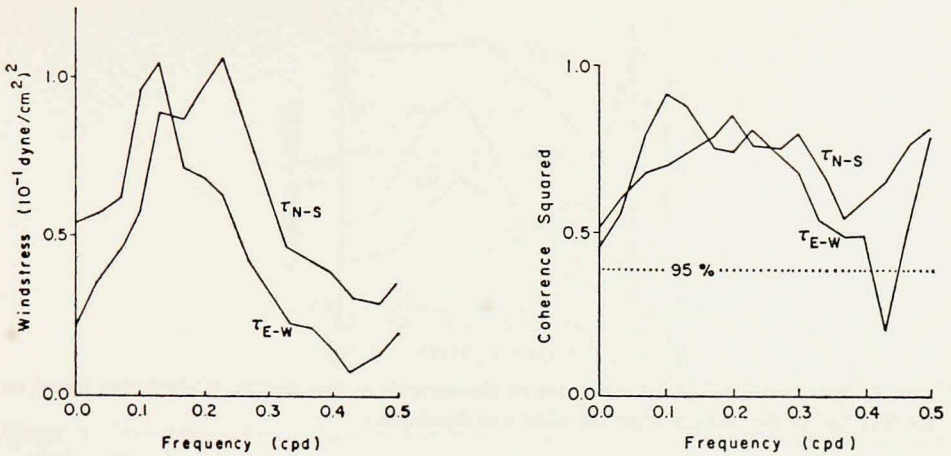
Figure 3. (a) Spectrum density of the N-S and E-W components of the Patuxent wind stress (Frequency is in cycles per day).

(b) Coherence between the Patuxent and Baltimore wind records for the N-S and E-W components. The dotted line represents the 95% significance level. (Calculation based on the total record length.)

shows the spectra of the N-S and E-W components of the wind stress at the Patuxent site. Most of the wind energy was contained at time scales of 3-10 days, and the N-S and E-W components were of approximately equal strength. Moreover, the winds were coherent between Baltimore and Patuxent at all frequencies (Fig. 3b); in general the Patuxent wind had a stronger E-W component while the Baltimore wind was stronger in the N-S direction. Due to the similarity between the two winds the Patuxent data are used in the following analysis; the general coherence of the low-pass wind agrees with previous results (Wang and Elliott, 1978) and suggests that the Patuxent wind can be taken as being representative of the wind distribution over most of the Bay and, in particular, of the wind at the mooring location.

The spatial structure of the nontidal currents. Both of the current records were dominated by the effects of the tropical storm; simple calculation showed that more than 60% of the current variance was contributed by the peak flow and Figure 2 demonstrates that the response to the river flow was highly coherent at the two depths. During the study period 75% of the total current energy was contained at time scales longer than 3 days. Linear regression, applied to the total records, gave that

$$U_{15} = 0.86 \, U_7 - 3.5 \, .$$

During the peak flow the maximum values of U_7 and U_{15} were 64.1 cm s^{-1}, and 52.3 cm s^{-1}, respectively. Therefore the observed ratio of U_{15}/U_7 was 0.82 while linear regression gave a ratio of 0.81.
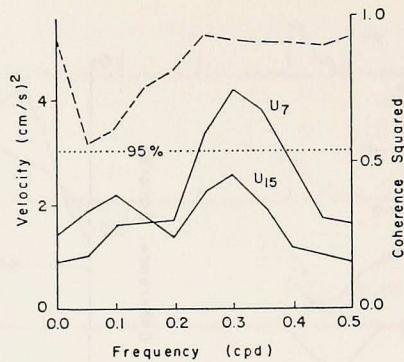
Figure 4. Spectrum density and coherence of the currents at two depths. (Calculation based on the first half of the records when the wind was dominant.)

Figure 4 shows the spectra and coherence of the currents during the first part of the experiment when wind forcing was dominant. Both spectra show a peak around 3 days and at this time scale the flow was vertically coherent and in phase. At time scales of 3-5 days the response function between U_7 and U_{15} had a magnitude of 0.82, U_7 leading U_{15} by 1-3 hours.

The response was barotropic at all time scales: a result which is in contrast to those reported by Elliott (1978) and Weisberg (1976) who found that the wind-driven flow near the mouth of an estuary can at times be highly baroclinic. Since the Potomac estuary was found to respond baroclinically to local wind forcing and barotropically to nonlocal effects this suggests that the Upper Bay is strongly influenced by nonlocal forcing. Since the two current records were highly coherent, and in phase, only the nearsurface flow will be considered throughout the following discussion.

In an analysis of lake dynamics Csanady (1973) showed that variations in cross-stream bottom topography can lead to the wind-driven generation of a coastal jet. Since the depth varies laterally across the Upper Bay it is possible that similar effects may have influenced the currents at the mooring position. In order to show that such a phenomena as proposed by Csanady does not account for the observations described here, it is necessary to demonstrate that the observed currents were representative of the laterally averaged longitudinal flow.

Knowing the geometry of the Upper Bay, we can estimate the sectional mean-flow, U, by using a relationship of the form

$$ U = \frac{Q}{A_c} - \frac{A_s}{A_c} \frac{\partial \eta}{\partial t}, $$

where Q is the river flow, η is the mean sea level upstream of the mooring, and A_c and A_s are the appropriate cross-sectional and surface areas.
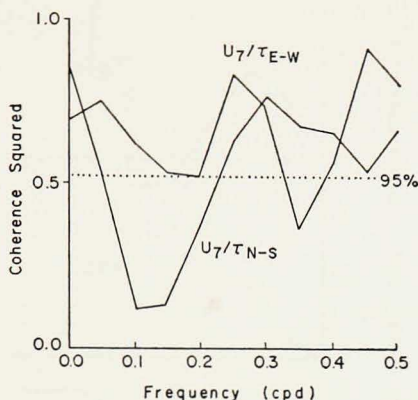
This leads to the relationship

Figure 5. Coherence between the E-W and N-S components of wind stress and the nearsurface current. The dotted line represents the 95% significance level. (Calculation based on the first half of the records.)

$$U = 0.43 \times 10^{-2} Q - 0.14 \times 10^5 \frac{\partial \eta}{\partial t}$$

where Q is in $m^3 s^{-1}$, and U and $\frac{\partial \eta}{\partial t}$ are in cm s^{-1}. The first term represents the flow due to river discharge and the second term represents the rate of change of the upstream volume. During the first part of the experiment the second term was dominant, while the first term was important during the peak flow. Therefore the two halves of the records were analyzed separately and the following results were obtained:

From linear regression of U_7 against river flow, we find that

$$U_7 = 0.46 \times 10^{-2} Q .$$

Alternatively, during the first part of the experiment when river flow could be neglected, a response analysis gave that

$$U_7 = - 0.15 \times 10^5 \frac{\partial \eta}{\partial t} .$$

Both terms are in good agreement with the expected result, and this suggests that the measured currents can be regarded as being representative of the sectionally averaged flow in this region of the Bay.

The influence of wind stress on the net flow. Figure 5 shows the coherence between the nearsurface flow and the E-W and N-S components of the wind stress. The wind forcing was most pronounced at time scales shorter than 5 days, however the E-W component was also significant at periods longer than 10 days. The calculations were made for other wind directions and the highest coherence between the current
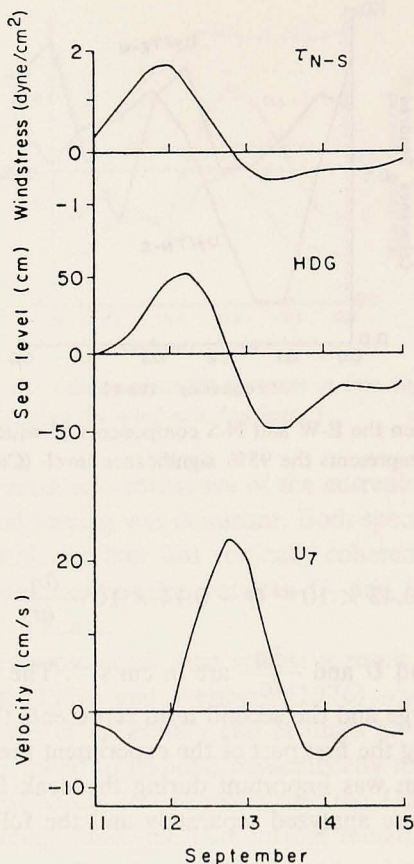
Figure 6. The time series of the N-S wind stress, sea level at Havre de Grace, and net current during the wind event of September 11-14.

and wind was found for a wind directed along a bearing between E and NE and at time scales of 2-3 days. These winds were approximately in phase with the current, i.e. an eastward wind caused seaward flow; this agrees with Wang and Elliott (1978) which showed that this effect is due to an Ekman transport in the Mid-Bay. Since the local river axis lies along a bearing of 40°, a downstream (seaward) flow appeared to be driven by an upstream wind; this was noted by Elliott and Hendrix (1976) and is an effect of the nonlocal Ekman response.

Figure 2 suggests that a southward wind resulted in a simultaneous positive (i.e. seaward) flow; in fact this is not correct. The northward wind led the current by 90°; at the 3 day time scale this corresponded to a lag of 18 hours. An example of this phase lag is shown in detail for the wind event on September 12. Figure 6 presents the N-S component of wind stress, the nearsurface flow and sea surface elevations for the period of September 11-14. The northward wind stress started increas-
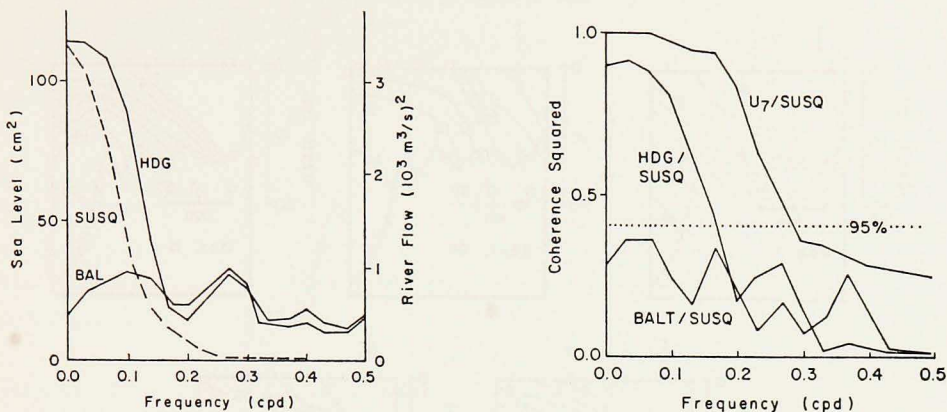
Figure 7. (a) Spectrum density of the Susquehanna discharge and the sea level at Havre de Grace and Baltimore Harbor.
(b) Coherence between the river discharge and the net flow, sea level at Havre de Grace, and sea level at Baltimore (Calculation based on the total record length.)

ing on September 11 and raised sea level throughout the Mid and Upper Bay. This was accompanied by landward (negative) net flow. The wind reached its peak at 0000 hrs on September 12 and the elevations peaked about 6 hours later; the net flow was near zero when the wind was at its strongest, implying that storage in the Upper Bay was then near its maximum. During September 12, as the northward wind relaxed, sea levels started to decrease and the current in the Upper Bay flowed seaward. The rate of decrease in sea level was greatest around 1800 hrs on September 12 and this resulted in a peak seaward flow; at this point the wind stress was near zero. As the wind stress increased in a southward direction the elevations continued falling, reaching minimum levels at 1200 hrs on September 13. At this time the seaward flow had decreased and reversed its direction, flowing landward again after 1200 hrs as the sea levels started to rise. The peak flow occurred 18 hours after the wind had reached its maximum and at a time when the northward wind was near zero, this agreed with the expected 90° phase relation.

The influence of river discharge. Since the Susquehanna supplies more than 95% of the fresh water to this region of the Bay, we may expect the river discharge to have a significant influence upon the nontidal circulation. Figure 7a shows the spectrum of the river discharge and also the spectra for sea level at Havre de Grace and Baltimore. The river flow spectrum was dominated by the effects of Eloise and most of the energy was contained at time scales longer than 10 days; there was negligible energy at time scales shorter than 5 days. The nontidal flow at the mooring location was coherent with the river discharge at time scales longer than 3 days (Fig. 7b) and lagged the river flow by about 10 hours. Figures 5 and 7 suggest that the net flow was driven by the wind and river discharge at separate time scales: the
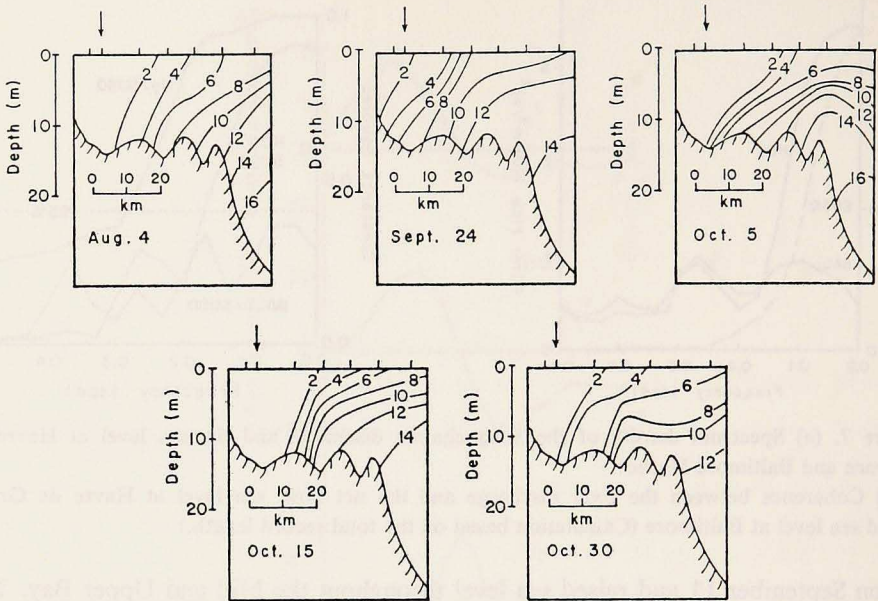
Figure 8. Salinity sections (‰) along part of the Upper Bay during the period of the measurements. The mooring position is marked by an arrow.

wind was important at periods shorter than 5 days while the river flow was the dominant mechanism at longer periods.

The spectrum of the Havre de Grace sea level was similar to the river flow and the two records were coherent and in phase at periods longer than 5 days, i.e. at time scales for which the river flow was significant (Figs. 7a and 7b). In contrast, the Baltimore sea level was not coherent with the river flow so that the influence of the Susquehanna on sea levels was confined to the upper 50 km of the Bay. Both Havre de Grace and Baltimore sea level displayed spectrum peaks around 4 days, probably due to wind forcing, and also around 2.5 days which is the period of the Bay's longitudinal seiche motion.

The influence of salinity. During the first part of the observation, prior to Tropical Storm Eloise, the overall mean surface flow was seaward at 1.6 cm s^{-1} while the deeper flow was landward at $-$ 2.7 cm s^{-1}. This suggests that salinity effects were driving an estuarine circulation at the mooring position. Figure 8 shows a series of salinity sections made along the Bay axis during the period of the measurements. On August 4 the water column was well-mixed at the mooring site and salinities were near zero. In contrast, by September 24 salinities had risen to 2-6‰ and there was significant stratification. Since the estuarine flow was well developed at the beginning of the current observations (August 10, Fig. 2) this suggests that the water column at the mooring had stratified rapidly between August 4 and August 10, and that
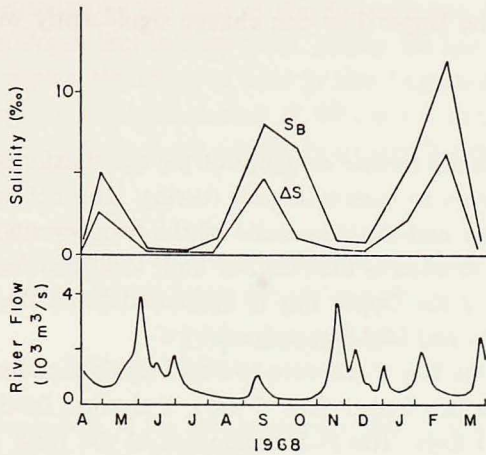
Figure 9. Bottom salinity and top-to-bottom salinity difference at the mooring location, and the daily mean values of Susquehanna discharge during the one-year period from April 1968 to March 1969. The salinity data were obtained during a series of monthly 'slack' runs.

the stratification had been maintained until the onset of Eloise. As shown by Figure 8, the peak river discharge following Eloise (September 27) caused a lowering of salinity of the Upper Bay, reducing salinities at the mooring to near zero. Salinity at the mooring remained insignificant for the remainder of the study and the estuarine circulation did not re-establish itself.

In the region between the mooring and a point 20 km downstream the bottom salinity varied by 10‰ (the highest gradients were observed on October 15), this produced a horizontal salinity gradient of about 0.5‰/km. In contrast, in the seaward portion of the Bay the salinity increased from about 10‰ to 30‰ at the Bay mouth over a distance of 250 km. This implied a mean gradient of 0.08‰/km, about an order of magnitude weaker than the Upper Bay salinity gradient.

The mean location of this front, which separates the fresh water on its landward side from the more saline stratified water on its seaward side, appears to be strongly dependent upon river flow. In Figure 9 is shown the bottom salinity at the mooring location and the top-to-bottom salinity difference plotted with the Susquehanna discharge for a one-year period during 1968-69 (Seitz, 1971). There is a clear visual correlation between the salinity and the river flow; the highest salinity and stratification was reached during periods when the river flow was low and, conversely, the salinity was reduced to near zero values following the peak flows of March and November of 1968 and late March of 1969. In late February, 1969, bottom salinity reached a maximum of 11.9‰. Then within the following weeks the water column reduced in salinity to values of less than 1‰; Figure 9 suggests that most of this reduction took place within a couple of days as a result of the peak flow which occurred during late March. It is therefore probable that the position of the front

and the salinity of the Upper Bay can change significantly with a time scale of a few days.

4. Discussion

The present results are further evidence of the complexity of the response of the Chesapeake Bay waters to meteorological forcing. The net currents depend upon both the directionality and the time scale of the wind events; in addition, the response at a selected location is the result of both local and far-field effects. In particular the response of the Upper Bay is influenced by nonlocal effects due to the response of the coastal and Mid-Bay waters.

Near the head of the Bay there were two distinct time scales: the river discharge was important at periods longer than 5 days, and wind forcing was important at time scales around 3 days. The N-S component of the wind generated a local response with a phase lag of about 6 hours between maximum wind and maximum elevation of the sea level; the maximum currents occurred 90° later, or about 18 hours after the peak wind. The E-W wind caused a nonlocal response due to an Ekman flux in the Mid-Bay and this resulted in an upstream wind generating a seaward net flow.

At all time scales the wind-driven response appeared to be barotropic. Baroclinic effects were only significant during the first part of the study when an estuarine circulation was established near the mooring position. The presence of an estuarine circulation was determined by the location of a front which separated low salinity homogeneous water at the head of the Bay from the significantly more saline and stratified water which lay to the south. Regions situated landward of the front have net seaward flow at all depths and the intensity of the flow is related to the strength of Susquehanna discharge. Salinity changes of 10‰ occurred over a separation of 20 km through the front and, as a result, the circulation on the seaward side of the front was influenced by density effects. The close correspondence between the location of the velocity stagnation point and the extent of the salinity intrusion has also been found in model studies (Festa and Hansen, 1976).

In a year-long study into the sedimentation of the Upper Bay, Schubel (1968) reported that the mean position of the turbidity maximum was located approximately 10 km downstream of the present study area and depended upon the river flow during the period of high runoff. The rate of deposition and resuspension within this region was found to be influenced mainly by the tide. However, the wind forcing also appears to be important since the wind-driven net currents can approach a significant fraction of the peak tidal velocities. If we assume a quadratic drag law, a peak tidal flow of 40 cm s^{-1} and a typical wind-driven flow to be 20 cm s^{-1}, then our estimate of the bottom stress will double when the two flows are in phase.

Haas (1977) has shown that the stratification in some estuaries can be related to the spring/neap cycle of the tides, however, in the Upper Bay the stratification is

more likely to be controlled by the flow of the Susquehanna which determines the mean position of the Upper Bay salinity front. Under the low river flow condition, typically in fall and winter, the effect of wind is also important at time scales of 3 to 5 days. For example, a wind-driven flow of 20 cm s^{-1} can induce a large horizontal displacement ($\sim$ 20 km) of the salinity front, and hence change in stratification, over a one-day period.

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