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Covariability of dissolved oxygen with physical processes in the summertime Chesapeake Bay

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ABSTRACT

Long, rapidly sampled time series measurements of dissolved oxygen, temperature, salinity, currents, winds, tides, and insolation were collected during the summer of 1987 across the mesohaline Chesapeake Bay. Analyses of the data show that short term variability of dissolved oxygen was both large and spatially heterogeneous. Time scales of variability ranged from the longest period fluctuations resolved (several days) to the sampling interval (several minutes). The largest variability was associated with large amplitude, wind and tide forced lateral internal oscillations of the pycnocline in the mainstem of the Bay. These resulted in advection of saline, hypoxic water from below the pycnocline onto the flanks of the Bay and into the lower reaches of the Choptank River, an adjoining tributary estuary. Advective variability at higher frequencies was likely due to internal waves, internal mixing, and/or spatial patchiness. Dissolved oxygen also responded to the daily cycle of insolation, but lagged insolation by at least 90° (6 h). Advective variability of dissolved oxygen is implicated as an important characteristic of the majority of summertime benthic environments in the mesohaline Chesapeake Bay and lower reaches of adjoining tributaries.

1. Introduction

Seasonal depletion of dissolved oxygen (DO) in subpycnocline waters is often observed in estuaries and the nearshore waters of continental shelves (e.g., May 1973, Falkowski *et al.*, 1980; Taft *et al.*, 1980; Officer *et al.*, 1984). This depletion generally results from a combination of biological and physical factors. In the Chesapeake Bay, for example, the annual spring freshet delivers large volumes of both fresh water and nutrients. In combination with increasing temperature and light, high nutrient levels result in a large increase in phytoplankton biomass that subsequently sinks out of the euphotic zone into subpycnocline waters where it is rapidly metabolized (e.g., Malone *et al.*, 1986; Tuttle *et al.*, 1987; Malone *et al.*, 1988). Net oxygen demand in subpycnocline waters is exacerbated by decreased mixing and aeration due to the increased stratification associated with the spring freshet and by the lower solubility of

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DO in seawater during the summer. The estuarine circulation further fuels subpycnocline oxygen demand by increasing the residence time of suspended particulate material within the system. The result is extensive hypoxia and, in recent history at least, anoxia in the subpycnocline waters of the Chesapeake during the summertime (e.g., Taft *et al.*, 1980; Officer *et al.*, 1984). Interannual variability in summertime DO levels is naturally large and confounds efforts to identify anthropogenic effects (Seliger and Boggs, 1988; Malone *et al.*, 1988).

Hypoxia and/or anoxia are not continuous throughout the summertime, however. Short term variability of DO may result from physical, biological and/or chemical processes. Observed short term variability of DO often is attributed to physical advection and reaeration-mixing events (Carter *et al.*, 1978; Malone *et al.*, 1986; Magnien, 1987; Breitburg, 1990). On the other hand, biological and chemical oxygen demand can produce rapid declines in DO (Kemp *et al.*, 1987, Malone *et al.*, 1988), consuming any DO available in subpycnocline waters in a matter of days. The diel cycle of production and respiration in the euphotic zone should cause a diel periodicity in near-surface DO (Tijggen, 1979). Phytoplankton blooms should result in even more dramatic increases and decreases in DO. The processes that produce short term variability in DO can have important ecological consequences (e.g., Malone *et al.*, 1986; Magnien, 1987; Breitburg, 1988). In addition, short term variability may be a source of error for measurements of longer term variability.

The relative contributions of physical, biological, and chemical processes to short term variability of DO are probably quite different from their contributions to seasonal and interannual variability. Biological and chemical rate processes often are regarded as steady over short time scales. Physically, advection may be more important to short term variability than the delivery of nutrients and oxidisable substrates and the maintenance of stratification that control seasonal fluctuations. At present there is little understanding of the dynamics of short term variability of DO, partly due to the limited technologies available for making long, rapidly sampled field measurements. Without these data, characterizations and explanations of short term biological and chemical variability in the marine environment are incomplete.

2. Methods

The data presented here represent the results of three research programs carried out during the summer of 1987 in the mesohaline portion of the Chesapeake Bay and the lower reaches of the Choptank River, an eastern short tributary estuary of the Chesapeake (Fig. 1). In one program, a near bottom mooring was maintained in relatively shallow water near the western shore of the Bay, in an effort to understand the impact of hypoxic intrusions on communities of epibenthic fishes. In another program, four moorings were maintained across the axis of the Bay and into the Choptank River in an effort to understand mechanisms controlling intrusions of hypoxic water from the lower layer of the Bay into the lower Choptank. In a third

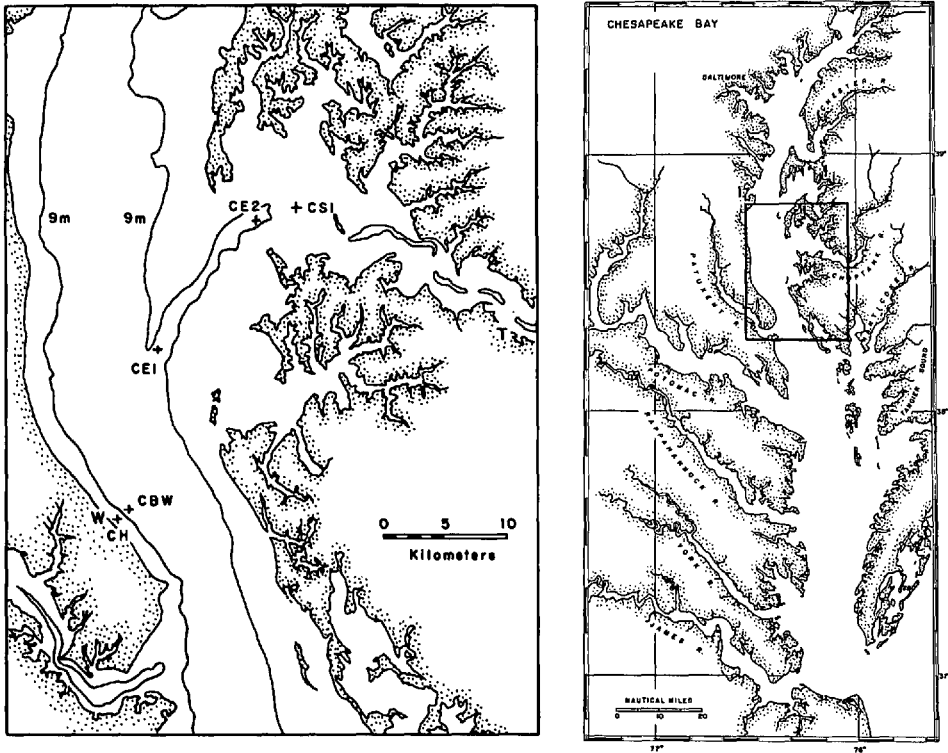


Figure 1. (a) Map of the study area, showing mooring locations and other reference points in the text. "W" denotes the location of the Calvert Cliffs wind station and "T" (center right) denotes the location of the Cambridge tide gauge. Site names signify: CH—Chesapeake Hydrolabs, CBW—Chesapeake Bay West; CE1—Choptank Entrance 1; CE2—Choptank Entrance 2; CS1—Choptank Sill 1. (b) Map of the Chesapeake Bay, showing the location of the study area in (a). "I's denote the locations of the insolation stations.

program, DO sensors were added to two of the moorings of the second program in order both to test a new technology for making remote moored measurements of DO in an estuarine environment, and to better understand the dynamics and consequences of DO variability in the mainstem Chesapeake Bay. Mooring locations are shown in plan view in Figure 1 and in cross-section in Figure 2.

Data from the shallow western shore mooring consist of time series of salinity, temperature, and DO sampled and recorded at 15 min intervals at a height of 10–20 cm above the bottom (3.7 m depth). The data were obtained with two Hydrolab 2040 Datasondes fitted with recessed cathode dissolved oxygen sensors. Each Datasonde was deployed for a 3–4 d interval before being replaced by a cleaned and recalibrated instrument, allowing a continuous time series to be constructed. Supplemental point measurements of salinity and DO were obtained during some deployments and recoveries with a Beckman RS5-3 field salinometer and a YSI 57 DO meter,

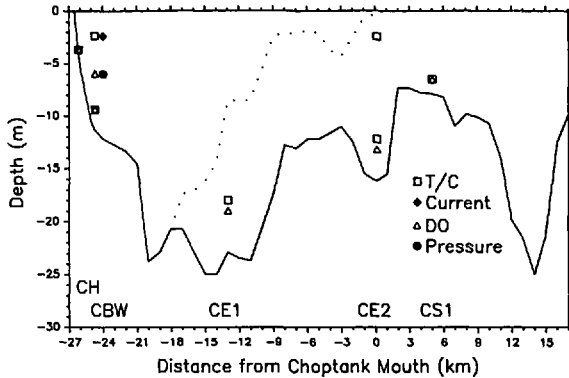


Figure 2. Elements of the moored arrays referred to in the text, along the transect of the mooring locations in Figure 1 and the axis of the Choptank River. The dotted line shows the minimum depth of the bar bounding the relict Choptank entrance channel on the north side. Km 0 is at the mouth of the Choptank River.

each calibrated independently. The average magnitude of the difference between DO measured with the Datasondes and the YSI meters was $0.4 \pm 0.1 \text{ mg L}^{-1}$; important fluctuations in measured DO described below are much greater than 0.4 mg L^{-1} .

Data were collected from the larger moored array using a variety of instruments. DO times series were measured using Endeco 1184 and 1133 Pulsed DO Sensors. Pulsed DO sensors use a standard polarographic DO probe but determine DO by sampling the transient response of the sensor to sudden excitation rather than the steady state current resulting from a continuously applied voltage (Langdon, 1984). Individual probe tip-membrane combinations were deployed sequentially in the same instrument body in the field, switched out at intervals of 1–4 weeks. An oxygen permeable anti-fouling wax was applied to the center of the membrane for the longer deployments. Laboratory calibrations were performed both prior to deployment and following recovery of individual probe tips. Bottle samples for determination of DO by Winkler titration (Carpenter, 1965) were taken during DO sensor redeployments, and supplemental DO information was obtained during semiweekly hydrographic surveys of the lower Choptank using Hydrolab Surveyor IIs. Final calibration procedure involved correction of the laboratory derived calibration coefficients for each probe tip-membrane combination using *in situ* point observations, insisting on certain reasonability criteria ($\text{DO} > 0$ and most observations $<$ saturation) where necessary. Based on the difference between one times series and six nearby point observations, pulsed DO probes can be accurate to approximately 0.1 mg L^{-1} over at least a 28 d deployment. The instruments appeared quite resistant to the effects of bio-fouling. They were affected adversely by hydrogen sulfide poisoning (Hale, 1983) although the effect was much slower than with a steady state DO sensor (W.R. Boynton, personal communication).

Conductivity, temperature, and current time series were measured with Endeco SSM current meters at sites CBW, CE1, and CE2, and with an InterOcean S4 at site CS1. Pressure time series were measured at mid-depth using Sea Data CTDs. Supplemental point measurements of conductivity and temperature were obtained during Choptank hydrographic surveys and also during current meter redeployments using a Chesapeake Bay Institute Induction Conductivity Temperature Indicator ("ICTI"; Schiemer and Pritchard, 1961).

Figure 2 shows the elements of the moored array selected for the present analysis. A 28 d period from August 12, 1987 to September 9, 1987, was chosen because DO and salinity time series were nearly continuous during this interval and apparently were not severely affected by either fouling or H_2S poisoning. Current data return was not complete during this interval but the near surface (2.4 m) record from mooring CBW on the western shelf of the Bay was available as a reasonable representation of the current response of the mainstem of the Bay. Salinity data from 6 m depth at this same mooring (one of the DO sensor locations) was not available, but was replaced by salinity measured at 2.4 m as the closest proxy. Pressure at mooring CBW 6 m was used to represent tidal height. Although this pressure record was not corrected for barometric pressure and steric height changes, direct comparison to the Cambridge tidal height record (NOS gauge) showed little difference in amplitude. The phase of the tide measured at CBW was, however, much more central to the moored array than the phase of the tide measured at Cambridge.

Additional hydrographic and environmental data were obtained from a number of sources. Temperature, salinity, and DO distributions from nearby stations in the mid-Bay were obtained from a Maryland Sea Grant sponsored program (M.R. Roman, unpubl. data). Chesapeake Bay hydrographic data during August, 1987, were obtained from the USEPA Chesapeake Bay Program Office. Hourly total irradiance time series data were obtained from the Maryland Agricultural Research Station in Wye, MD (R. Brinsfield, unpubl. data), and daily integrated total irradiance and photosynthetically active radiation (PAR) data were obtained from the Smithsonian Environmental Research Center in Edgewater, MD (B. Drake, unpubl. data). The latter were used to convert the total irradiance time series data to a time series of PAR by a factor derived from least squares regression ($r = 1.00$):

$$PAR (E m^{-2} h^{-1}) = 0.0924 \times TRAD (\text{langleys } h^{-1}) \quad (1)$$

Hourly wind data were obtained from the Baltimore Gas and Electric Power Plant at Calvert Cliffs (M. Wieland, unpubl. data). Tidal height data (in addition to the pressure data obtained from the moored instruments) were obtained from the NOS gauge at Cambridge, MD. Wind, tide, and insolation stations are shown in Figure 1.

All salinity data are presented in practical salinity units (psu; Lewis, 1980). One psu is essentially equivalent to 1 ppt, but salinity in psu is referenced to a standard and is dimensionless.

Multiple input cross-spectral analysis (Jenkins and Watts, 1968; Bendat and Piersol, 1971, Brillinger, 1975) was used to examine relationships between the variability of DO, insolation, salinity, etc. as a function of frequency. All time series were aligned, smoothed by hourly averaging, and resampled at hourly intervals. A 512 h period starting at 0000 on August 13, 1987 (21.33 d) was chosen as the largest possible common time period with power of 2 length. Frequency band averaging was used to improve the stability of the spectral estimates, with a minimum of 9 adjacent spectral estimates averaged together. As a result of these constraints, the lowest frequency resolved was 0.234 cpd (4.26 d period) and the highest frequency was 10.27 cpd (2.34 h period). Spectral analysis was performed in two parts for the data from station CH 3.7 m because the time series of salinity and DO were not continuous. As a result, the spectra from CH are slightly less well resolved than their counterparts from other stations.

3. Observations and analyses of variability

a. The seasonal and spatial context. In the late summer, freshwater inflow to the Chesapeake Bay is generally at or near its annual minimum and seasonal anoxia in the lower layer of the mainstem has been established for some time (e.g., Malone *et al.*, 1988). In August, 1987, mid-Bay longitudinal gradients of salinity and DO in the study area were weak or nonexistent (Fig. 3a and 3c). Vertical gradients of salinity and DO were much greater than the longitudinal gradients, especially through the pycnocline. Lateral gradients in salinity and DO were due primarily to the lateral tilt of the pycnocline (Fig. 3b and 3d). In the mid-Bay region of interest, low DO corresponded to high salinity below the pycnocline, high DO corresponded to low salinity above the pycnocline, and the isohaline and iso-oxygen surfaces tended to coincide; i.e., there was a strong inverse spatial correlation between salinity and DO.

b. Observed variability of DO and salinity. The observed time series of DO and salinity for the period selected are presented in Figure 4, along with their respective means and standard deviations. [Note that the apparent nonzero floor in portions of the records from CBW 9.4 m and CE1 19.0 m is an artifact caused by instrument malfunction.] Average DO concentrations were at an absolute minimum in the deep center channel of the Bay and increased both with distance away from the channel and with relative proximity to the surface at any given location. The pattern of average salinity was inverse to that of DO. Measured temporal variability in both DO and salinity was as large as spatial variability at the study site (Fig. 3). Measured standard deviations of DO over a one month period were, on average, almost 50% of their respective means. Maximum DO variability occurred near the depth of the mainstem Bay pycnocline (stations CBW 9.4 m and CE2 13.1 m). The spatial pattern of salinity

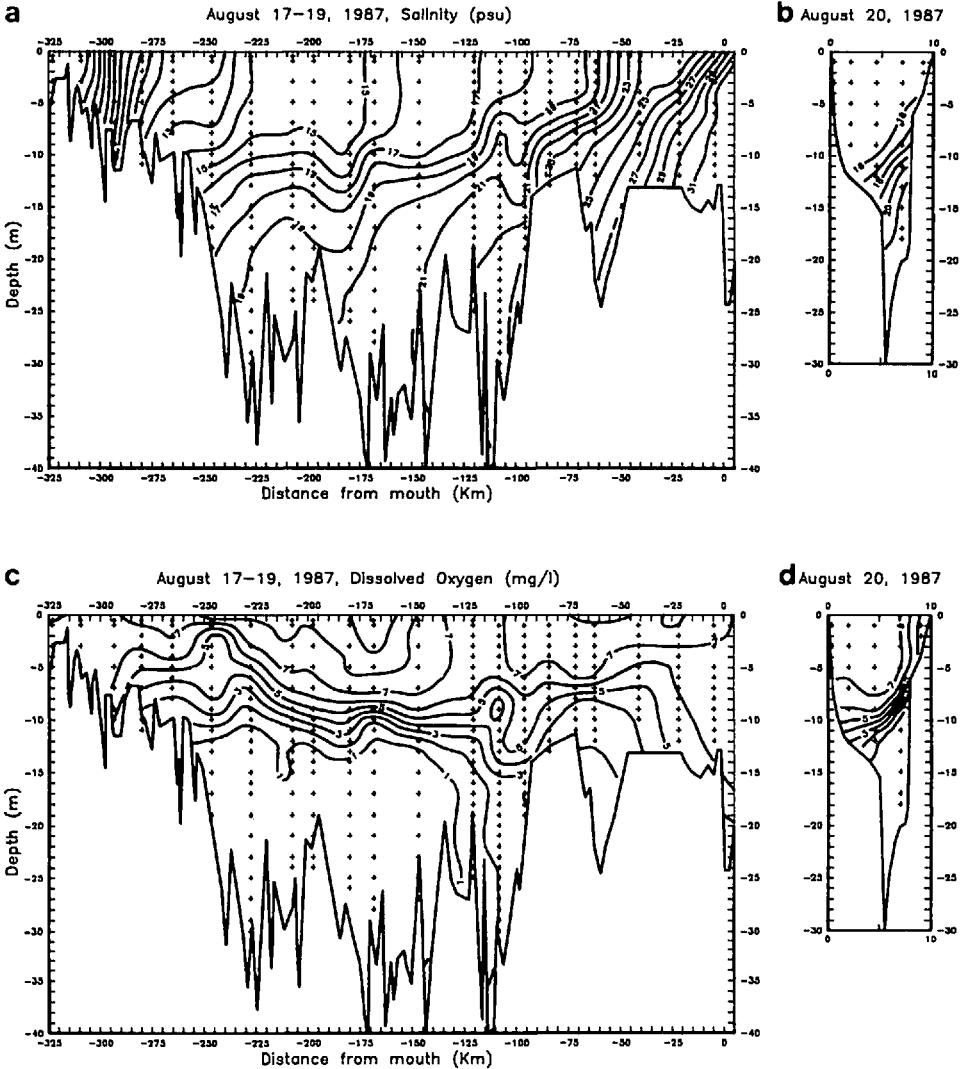


Figure 3. Hydrography of the Chesapeake Bay, August 17-20, 1987. (a) Salinity distribution along the axis of the Bay. Mainstem station locations for the present study were between km -185 and km -197. Vertical exaggeration of 5000. (b) Salinity distribution across the axis of the Bay at km -192 of (a). Vertical exaggeration of 1000. (c) Axial DO distribution. (d) Cross-axis DO distribution. Axial data from USEPA Chesapeake Bay Program monitoring data base. Cross-axis from MD Sea Grant sponsored program, courtesy of M. Roman.

variability was similar to that of DO, except that salinity variability at the deep center channel station (CE1 19 m) was relatively greater.

Individual fluctuations in DO were often mirrored by inverse fluctuations in salinity, indicating advection of low DO from subpycnocline waters past the point of observa-

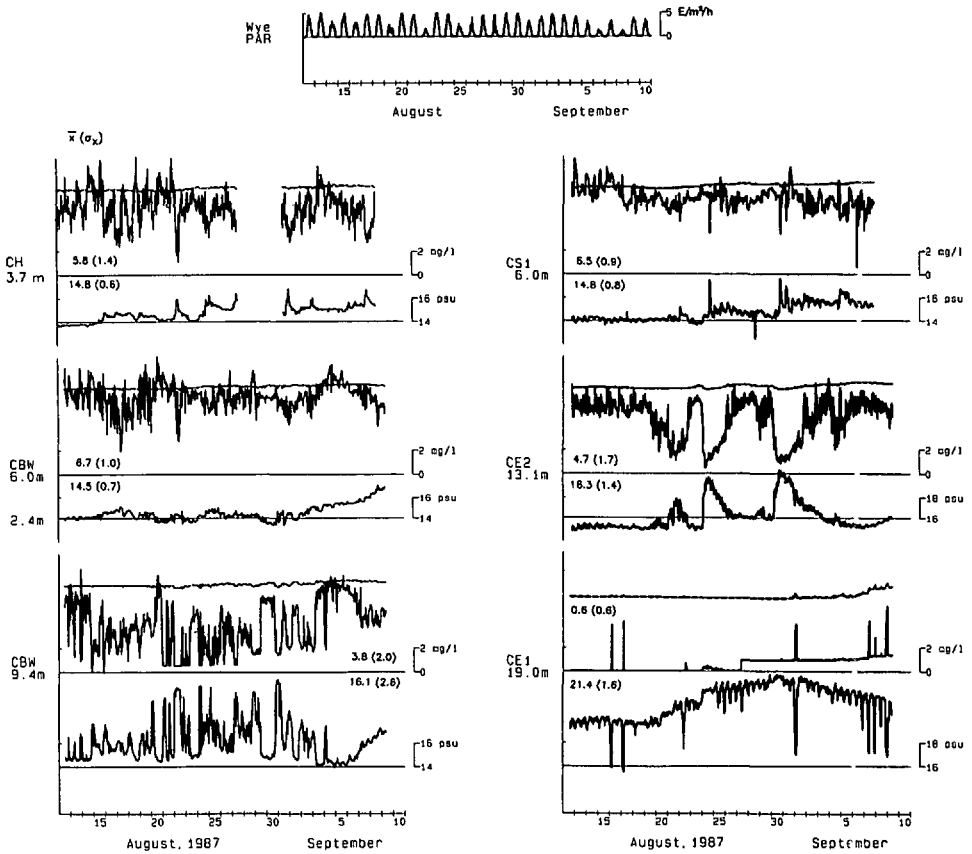


Figure 4. Time series of observed DO (solid line in upper plot of each panel), DO at saturation for measured temperature and salinity (dashed line in upper plot of each panel), observed salinity (lower plot of each panel), and insolation (separate plot). Mean (\bar{x}) and standard deviation (σ_x) for each of the observed time series of salinity and DO are also shown. Panels are arranged with the shallowest stations at the top and the western shore stations on the left.

tion. This is clear in the time series from the deeper stations and can be seen to a lesser extent in the larger spikes of low DO from the shallow stations. Time series from the different stations are quite distinct. For example, the DO time series from CBW 9.4 m shows large, rapid changes over relatively short time periods of hours to days, while that from CE1 19 m shows very little variability except for occasional spikes of high DO, and the time series from CE2 13.1 m shows large, rapid decreases in DO followed by slower increases with a dominant time scale of several days.

Temperature and salinity induced changes in the saturation point for DO explain little, if any, of the observed variability of DO. Saturation levels for DO varied slightly from station to station, but remained almost constant relative to observed DO variability.

c. *Cross-spectral analysis.* Optimal results were obtained from multiple input cross-spectral analysis at each station by subdividing the procedure into two stages. In the first stage (Fig. 5), DO was considered as the single output with salinity and insolation as inputs. In the second stage (Table 1), salinity was considered as the output with wind and tidal height as inputs.

Selected results of the DO-salinity-insolation analysis are presented in Figure 5 as: (1) the energy density (frequency distribution of variance) of all three time series; (2) the partial coherence (frequency distribution of partial correlation squared) between each input and DO; (3) the multiple coherence (frequency distribution of multiple correlation squared) between both inputs and DO; and (4) the phase of the transfer function (frequency distribution of the phase lag) between each input and DO. Phase estimates are presented only when they are significant at the 90% confidence level. High partial coherence between salinity and DO with a $\pm 180^\circ$ phase lag indicates the importance of advection of saline, low DO water past the measurement point. High partial coherence between insolation and DO indicates a direct response of phytoplankton productivity to incident light. Local production might be expected to result in a near-zero to 90° phase lag between PAR and DO in the euphotic zone, depending on the balance between photosynthetic oxygen production and heterotrophic demand. Spectra from the second part of the CH 3.7 m record are not presented in Figure 5 because they are very similar to spectra from the first part of the record, and spectra from CE1 19 m are not presented because the DO time series is too irregular for spectral decomposition of the variance to be meaningful.

The energy density spectra of DO, salinity, and PAR help to quantify the similarities and differences observed in the raw time series presented in Figure 4. The spectrum of PAR in each panel (except for CH 3.7 m) is constant, so that it provides a reference level for comparison between figures. The spectrum of PAR from CH 3.7 m represents a shortened segment of the PAR time series corresponding to the first DO and salinity segment in Figure 4, but there is no significant difference between this spectrum and the constant PAR spectrum of the other panels except for decreased resolution. As expected, the primary peak in the spectrum of PAR is at 1 cycle per day (cpd). The significant secondary peak at 2 cpd reflects the nonsinusoidal daily variation of light: insolation was roughly sinusoidal during the day but constant at 0 during the night, which caused harmonics of the fundamental diurnal frequency to be introduced into the spectrum.

In general, the spectra of DO and salinity from a given location have roughly the same shape. Energy density decreases with increasing frequency more steeply than the -1 slope corresponding to a constant variance preserving spectrum, except for isolated peaks at 1, 2, and 4 cpd. In other words, the total variance of both DO and salinity tended to decrease with increasing frequency, with some exceptions at 1, 2, and 4 cpd. With the exception of the spectra from CH 3.7 m, all of the DO and salinity spectra show clear peaks at semi-diurnal frequency (2 cpd). In the case of CH 3.7 m, the

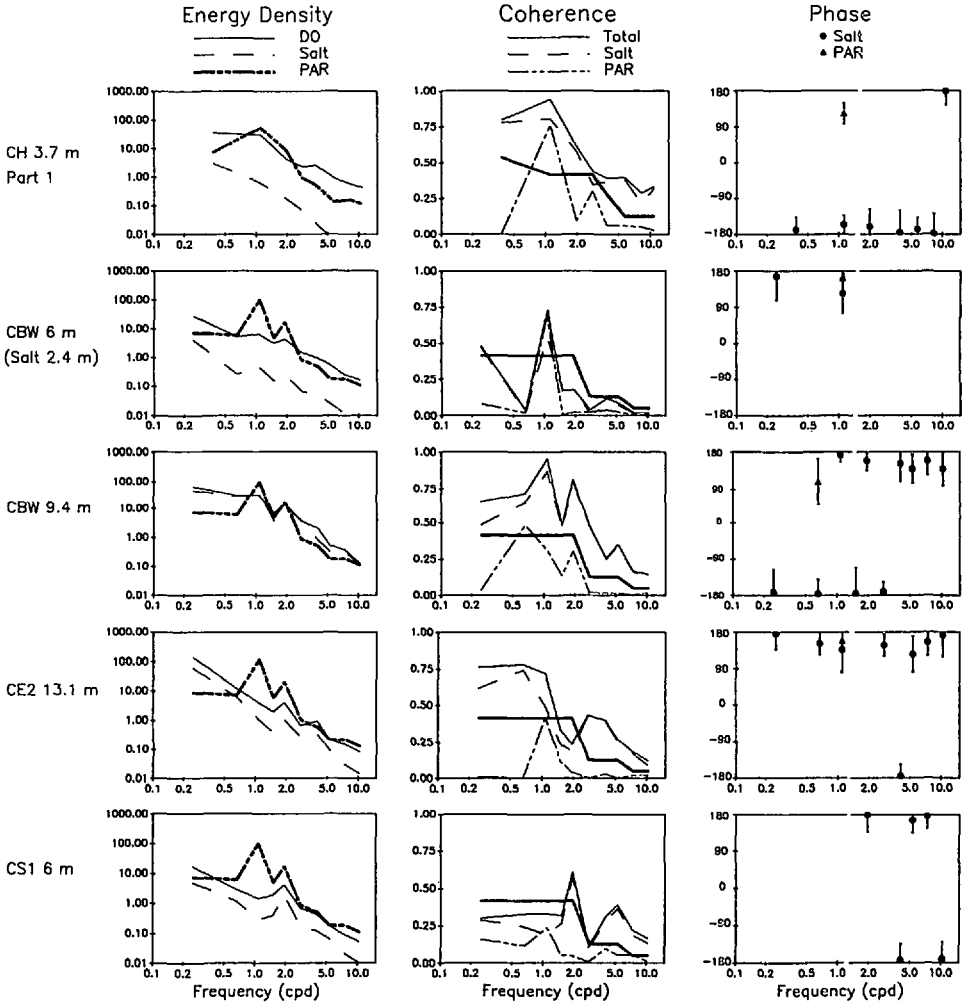


Figure 5. Selected results of DO-salinity-insolation cross-spectral analysis. *Left column*—energy density spectra. *Center column*—multiple coherence (“total”), salinity-DO partial coherence, and PAR-DO partial coherence. Heavy solid line shows 90% confidence level for significant coherence. *Right column*—salinity-DO phase and PAR-DO phase. Only phase values that are significant at the 90% confidence level are plotted, with error bars. Station names are shown to the left of their corresponding rows.

resolution is not sufficient to pick out a semi-diurnal peak, but a semi-diurnal peak is evident in spectra with greater resolution. Regular semidiurnal variability in salinity was also common at CE1 19 m, with occasional semi-diurnal variability in DO (Fig. 4). Thus, semi-diurnal variability was ubiquitous. Most of the DO and salinity spectra from the main Bay stations also show a clear peak at diurnal frequency (1 cpd). This diurnal peak is conspicuously absent in the spectra from within the lower Choptank

Table 1. Partial coherences and phases between CBW tidal height and salinity measured at each station. Values derived from multi-component cross-spectral analysis with salinity as the output and with wind and tidal height as inputs. Subtidal means the frequency band below diurnal, centered on 0.66 cpd except for the CH records. The 90% confidence limit for nonzero coherence is 0.459; only significant values are listed. "*" indicates that there was also significant partial coherence between N-S wind and salinity.

Station	Subtidal		Diurnal		Semi-diurnal	
	Coherence	Phase	Coherence	Phase	Coherence	Phase
CH 3.7 m	—*	—	0.50	39°	0.47	-167°
CBW 2.4 m	0.46	-1°	0.50	34°	0.87	77°
CBW 9.4 m	0.61*	3°	0.61	3°	0.78	86°
CE1 19.1 m	—	—	0.73	-145°	0.79	34°
CE2 13.1 m	0.48	7°	—	—	—	—
CS1 6 m	0.52*	18°	—	—	0.63	59°

River (CE2 13.1 m and CS1 6 m). The highest low frequency energy levels in both salinity and DO appear in the spectra from CBW 9.4 m and CE2 13.1 m, in agreement with Figure 4.

There are three main points to be made from the coherence and phase spectra:

1. Of the two inputs considered in the present analysis, advection of high salinity and low DO water past the point of observation clearly dominated observed variability of DO. The multiple coherence is almost entirely dominated by the partial coherence between salinity and DO, and 28 of the 32 significant phase estimates between salinity and DO are within 90% confidence limits of $\pm 180^\circ$.
2. Insolation contributed significantly to the observed variability of DO at 4 out of the 5 stations in Figure 5. Peaks in insolation-DO partial coherence are between 0.5–2 cpd. Significant phase estimates are generally between 90° and 180° . The best model for the influence of insolation on the observed DO variability therefore seems to be one of a limited diurnal response, with insolation leading DO by 90° or greater.
3. The influence of advection tended to increase with proximity to the bottom and with proximity to the center channel of the mainstem of the Chesapeake Bay, based on the magnitude and distributions of salinity-DO partial coherence estimates. Similarly, the influence of insolation increased toward the surface. Station CS1 6 m appears to be an exception: cross-spectra from CS1 show no significant insolation-DO coherence and the weakest salinity-DO coherence of the array (excluding data from CBW 6 m where the salinity and DO time series were not from the same depth).

d. Very high frequency variability. The highest frequency fluctuations resolved by the moored instrumentation (10–30 min) were not included in the spectral analysis because the hourly resampling required for spectral analysis does not resolve periods less than 2 h. Figure 6 illustrates high frequency variability, presenting a 10 h segment of high pass filtered (3 h half power point) DO and salinity time series from CBW

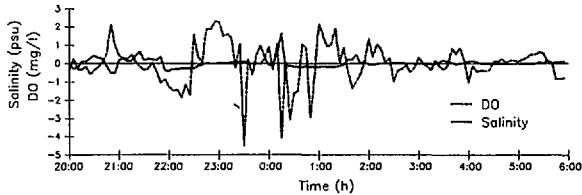


Figure 6. Very high frequency variability of DO and salinity, shown as a segment of high pass filtered (3 h half power point) DO and salinity time series data from CBW 9.4 m, on August 13-14, 1987.

9.4 m on August 13–14. The high frequency variability observed here was the greatest observed from the moored array, but similar high frequency variability was present in many of the records.

Variability at high frequencies was frequently large. There are eight instances in Figure 6 in which DO changed by 3–5 mg L⁻¹ in one sampling interval (5 min). Some of the DO fluctuations were matched by oppositely directed salinity fluctuations, indicating advection, but some were not. This kind of large high frequency variability tended to be episodic, occurring for short periods of time separated by long intervals with very little high frequency activity. Episodes of high frequency variability showed no apparent patterns of occurrence and there was no consistent relationship to lower frequency phenomena, indicating that a number of different processes were contributing to high frequency variability.

4. Processes responsible for the observed variability

The previous section showed that most of the observed temporal variability of DO and salinity was at subtidal and tidal frequencies, and was caused by the advection of inverse spatial patterns of DO and salinity past the station locations. At very high frequencies large DO and salinity variability was observed, but it tended to be episodic and the inverse relationship between DO and salt was not as clear. Insolation was also an important contributor to DO variability near 1 cpd, but the phase lag between insolation and DO was often larger than expected. This section explores the processes responsible for the several modes of variability, keeping always in mind that the net variability observed at any given time may have resulted from a multitude of different processes acting simultaneously. In examining the processes responsible for advective variability, the analysis is restricted to the behavior of salinity alone for the sake of simplicity.

a. Subtidal advective variability. In Figure 7, near-surface and near-bottom salinity records from moorings CBW and CE2 are plotted for comparison to Calvert Cliffs wind vectors, CBW tide records, and CBW longitudinal and lateral near-surface current records. The near-bottom salinity records from CBW and CE2 clearly show

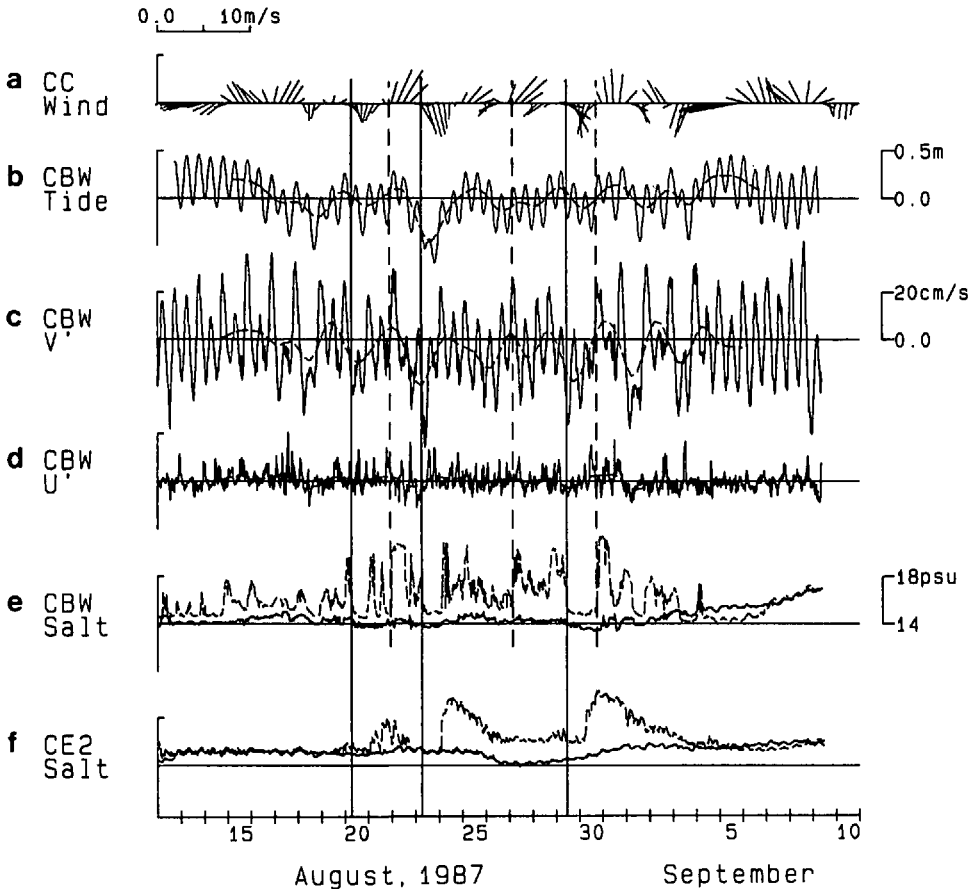


Figure 7. (a) Vector time series of subtidal (low pass filtered, 34 h half power point) Calvert Cliffs wind. (b) CBW tide, mean removed; solid line unfiltered, dashed line subtidal. (c) Longitudinal velocity, CBW 2.4 m, positive up-Bay; solid line unfiltered, dashed line subtidal. (d) Lateral velocity, CBW 2.4 m, positive to the east; solid line unfiltered, dashed line subtidal. (e) CBW salinity; dashed line 9.4 m, solid line 9.4 m, solid line 2.4 m, both unfiltered. (f) CE2 salinity; dashed line 13.1 m, solid line 2.4 m, both unfiltered. Solid vertical lines mark western shore downwelling episodes, dashed vertical lines mark western shore upwelling episodes.

the large fluctuations in salinity that were matched by oppositely directed DO fluctuations in Figure 4. The largest of these salinity fluctuations had characteristic time scales longer than either the semi-diurnal or diurnal tidal period, or subtidal characteristic frequencies. These subtidal salinity fluctuations were associated with changes in wind forcing. For example, the rapid decreases in CBW 9.4 m salinity marked by solid vertical lines in Figure 7 were coincident with episodes of southward wind and the rapid increases in CBW 9.4 m salinity marked by dashed vertical lines were coincident with episodes of northward wind. The physical mechanism that produced this association is not immediately obvious, however.

The strongest current response to wind forcing was longitudinal. In the three episodes of southward wind marked by the solid vertical lines, subtidal down-Bay (southward) surface currents lead southward wind by several hours to a day with maximum wind-driven current speeds of 10–20 cm s^{-1} . The longitudinal response was similar for episodes of northward wind, except that maximum wind-driven current speeds were only 5–10 cm s^{-1} . Subtidal sea level lagged slightly behind velocity. This longitudinal response agrees with the wind-driven surface layer response found by Pritchard and Vieira (1984) and Vieira (1986) in analyses of data from the same region of the Chesapeake Bay.

Advection of the longitudinal salinity gradient by wind-driven currents cannot explain the rapid response of the CBW 9.4 m salinity record, however. The longitudinal salinity gradient in the Bay (S_y) was about 0.04 psu km^{-1} (Fig. 3). The maximum longitudinal speed observed (V) during an episode of southward wind-driven flow was about 40 cm s^{-1} , adding the regular tidal currents to the wind-driven currents. The maximum rate of decrease in salinity that might have been due to longitudinal advection was therefore approximately $VS_y = 0.06 \text{ psu h}^{-1}$, which is two orders of magnitude slower than the corresponding observed rate of decrease in CBW bottom salinity of approximately 3 psu in 0.5 h or 6 psu h^{-1} .

The only salinity gradient strong enough to produce the observed rates and magnitudes of change in CBW bottom salinity was the vertical gradient through the pycnocline. For example, a vertical gradient of 0.5 psu m^{-1} through the pycnocline (Fig. 4) would require a vertical velocity of only 0.3 cm s^{-1} to produce a 6 psu h^{-1} rate of change. Furthermore, typical peak to peak salinity changes of 4–8 psu would require vertical excursions of only a few m, but would require longitudinal excursions of approximately 150 km. Thus, the large, rapid changes in salinity (and DO) observed at CBW 9.4 m must have been the result of vertical motion of the pycnocline through the moored array.

This vertical motion of the pycnocline was connected to wind forcing through lateral downwelling and upwelling that accompanied the longitudinal wind-driven response. Referring again to the three episodes of southward wind marked by the solid vertical lines in Figure 7, it is apparent that the downwelling response at mooring CBW was due to onshore advection in the surface layer that accompanied southward longitudinal flow. The salinity cross-section of Figure 3b was taken slightly to the north of the mooring transect on August 20, 1987, immediately following the first of the three western shore downwelling episodes identified in Figure 7. A downward excursion of approximately 10 m must have occurred to move the pycnocline from above CBW 9.4 m prior to downwelling to its position in Figure 3b. The corresponding offshore excursion of the benthic front at the intersection of the pycnocline and the sloping bottom must have been approximately 4 km.

The western shore downwelling response to southward wind was matched by an eastern shore upwelling response. This is shown indirectly by the salinity records from

CE2 in Figure 7. Sharp increases in salinity at CE2 13.1 m occurred about 17 h after the sharp decreases in salinity at CBW 9.4 m. These salinity increases marked internal surges of saline, low DO water that propagated up the entrance channel of the Choptank from the Bay and spilled over the sill at the mouth of the river, often reaching and passing site CS1 (Sanford and Boicourt, 1990). Entrance channel propagation times were nearly equal to the 17 h time lag in Figure 7. Thus, surges at CE2 are a delayed response to upwelling on the eastern shore that occurred simultaneously with downwelling on the western shore.

Episodes of northward wind produced the opposite lateral response; i.e., upwelling on the western shore accompanied by onshore intrusion of saline, low DO subpycnocline waters, and downwelling on the eastern shore. Western shore upwelling is apparent in the three episodes marked by vertical dashed lines in Figure 7. An eastern shore downwelling response is not apparent in the records from CE2, since the mouth of the Choptank is cut off from the Bay by a shallow sill, but the salinity record from CE1 19 m in Figure 4 shows that large negative spikes in salinity followed upwelling observed at CBW with some small lag. The lag may be explained by the fact that CE1 19 m was near the bottom of the vertical excursion of the pycnocline on the eastern shore. The response at CE1 indicates a downward vertical excursion of about 10 m from the position of the pycnocline in Figure 3b.

Upwelling and downwelling occurred in response to north-south (longitudinal) reversals in wind, rather than to onshore-offshore winds. This is in agreement with the numerical predictions of Chao (1988) and the observations of Breitburg (1990), but counter to the results of Carter *et al.* (1978), which were based on an analysis of data from a single site on the western flank of the Bay very close to site CH in this study. In the data presented here, cycles of north-south wind reversal occurred every 2–6 d. Sellner and Kachur (1987) estimated a similar low DO-high salinity periodicity (6.5 d) at an 11 m station off Calvert Cliffs, based on analysis of data from 1974–1983. The downwelling-upwelling response reported here is the same as the lateral “tilting” reported by Malone *et al.* (1986).

b. Diurnal and semi-diurnal tidal advection. Diurnal and semi-diurnal fluctuations were a prominent feature of all of the DO and salinity records from the mainstem Bay. Part of the diurnal variability in DO was in response to the daily cycle of insolation, but much of the diurnal variability in DO and most of the semi-diurnal variability in DO were due to advection.

In the mainstem of the Chesapeake Bay, the physical processes that produced this advective response were most likely internal tides. Diel wind fluctuations might also have been implicated, but the results of cross-spectral analysis between salinity, wind, and tidal height (Table 1) show no significant coherence between wind and salinity at 1 and 2 cpd. Tidal height was significantly coherent with salinity at both 1 and 2 cpd. Previous conclusions about the dominance of vertical advection hold equally for the

tidal fluctuations in Figure 4 and 7; i.e., rapid, large changes in DO and salinity at semi-diurnal and diurnal frequencies could only have been produced through vertical motion of the pycnocline. Vertical motion of the pycnocline at tidal frequencies that is phase locked with the diurnal and semi-diurnal surface tides indicates internal tides forced by the surface tides.

The data in Table 1 further suggest that this internal tidal response was a near-resonant lateral internal oscillation of the pycnocline. The phase difference between the stations that would be at the two opposite extremes of a first mode lateral internal oscillation (CH 3.7 m and CE1 19 m) was close to 180° at both diurnal and semi-diurnal frequencies; i.e., salinity was high at CH when it was low at CE1 and *vice versa*. In addition, there was a near 180° phase shift at stations CH and CE1 between the diurnal and the semi-diurnal frequency bands, which is consistent with the behavior of a dynamical system when crossing the resonant frequency.

This suggestion is supported by an estimate of the lowest mode natural internal period of this cross-section of the Bay. Referring to the cross-sectional salinity distribution shown in Figure 3, simplifying the complicated geometry of the Bay to a rectangular cross-sectional geometry, approximating the density distribution as surface and bottom mixed layers with a stratified interior, and estimating the first mode long internal wave speed, c_i , from the equations presented in Sanford and Grant (1987), we obtain $c_i \approx 0.3 \text{ m s}^{-1}$. The lowest mode natural period of the cross-section is then

$$T_1 = \frac{2W}{c_i} = \frac{2 \times 7.5 \times 10^3 \text{ m}}{0.3 \text{ m s}^{-1}} = 5 \times 10^4 \text{ s} = 14 \text{ h} \quad (2)$$

where W is the width of the model cross-section. This estimate is obviously rough, since it does not account for the actual geometry, probable nonlinearity, and friction, but it does show that a near-tidal natural period was quite likely. The calculation is relatively insensitive to the model geometry chosen, as long as the cross-sectional areas above and below the pycnocline are kept roughly constant and the total density difference from top to bottom is maintained.

The magnitude of the CBW tidal salinity oscillations on August 20–23 was less than the magnitude of the wind forced oscillations, suggesting that the internal tidal height was approximately 5 m on the flanks of the Bay. Thus, a tidal salinity or DO signal would only be present with the “average” position of the pycnocline was within 2–3 m of the depth of observation. If the internal tide were a lateral oscillation, its amplitude would be considerably less in the center of the Bay than on the flanks of the Bay. Finally, a longitudinal component of the internal tides cannot be ruled out. Longitudinal internal tides have been shown to be important in highly stratified estuaries (e.g., Jay and Smith, 1990) and there are numerous potential locations along the axis of the Bay for generation of internal tides by the interactions of surface tides and topography.

The lower Choptank River was effectively isolated from the internal tidal activity in the mainstem of the Bay, most probably because of the constrictions imposed by the geometry of the Choptank entrance channel. Spectral analysis showed no strong diurnal signal at CE2 13.1 m and CS1 6 m. Data in Table 1 further show that there was no significant coherence between tidal height and salinity in the diurnal band at either station, and none in the semi-diurnal band at CE2 13.1 m. Semi-diurnal tidal advection of high salinity and low DO at CS1 6 m was probably longitudinal advection of the benthic front produced by wind-forced intrusions into the Choptank. This front often lies in the vicinity of mooring CS1 (Sanford and Boicourt, 1990).

c. The diurnal insolation response. Cross-spectral analysis has shown a maximum response of DO to insolation at or near 1 cpd. This response was stronger near the surface. Both of these characteristics were expected. However, the observed phase lags between insolation and DO, which often exceeded 90°, were not entirely expected.

A phase lag of about 90° between insolation and DO is easily explained as the phase lag between net oxygen production plus inward surface flux (source terms) and net consumption plus loss (sink terms) in the surface mixed layer. DO increases during daylight hours as long as the sources are larger than the sinks, reaching a maximum just before sundown. DO decreases during the night when the sinks are larger than the sources, reaching a minimum at or slightly after sunrise. The daily maximum and minimum in DO are thus lagged roughly 90° behind the daily maximum and minimum in insolation. This mechanism is illustrated schematically in Figure 8, where sources and sinks other than production and respiration have been ignored.

The explanation for phase lags greater than 90° may have to do with time-dependent downward diffusion of DO from the surface layer towards the depth of measurement. The problem of heat conduction into a very thick wall subjected to periodic surface temperatures (e.g., Chapman, 1974) is analagous. Temperatures within the wall vary in time with the same periodicity as the driving surface temperatures, but with a phase lag that increases with distance away from the surface. The amplitude of variation decays exponentially away from the surface. Both the phase lag (in radians) and the decay factor are given by

$$\epsilon = z(\pi\omega/\alpha)^{0.5} \quad (3)$$

where z is the distance below the surface, ω is the radian frequency of the forcing, and α is the thermal diffusivity (*ibid.*). In the present case, z would be the distance below the base of the surface mixed layer, ω would be 1 cpd = $7.27 \times 10^{-5} \text{ s}^{-1}$, and α would be replaced by the turbulent diffusivity, K_z . For example, if we assume $K_z = 1 \text{ cm}^2 \text{ s}^{-1}$ at $z = 1 \text{ m}$ into the pycnocline, then the phase of DO relative to insolation should be lagged an additional 1.5 rad = 86° behind the 90° phase lag of the surface mixed layer and the amplitude of DO fluctuations should be decayed to $e^{-1.5} = 0.2$ of the amplitude in the surface mixed layer. This response is also illustrated in Figure 8.

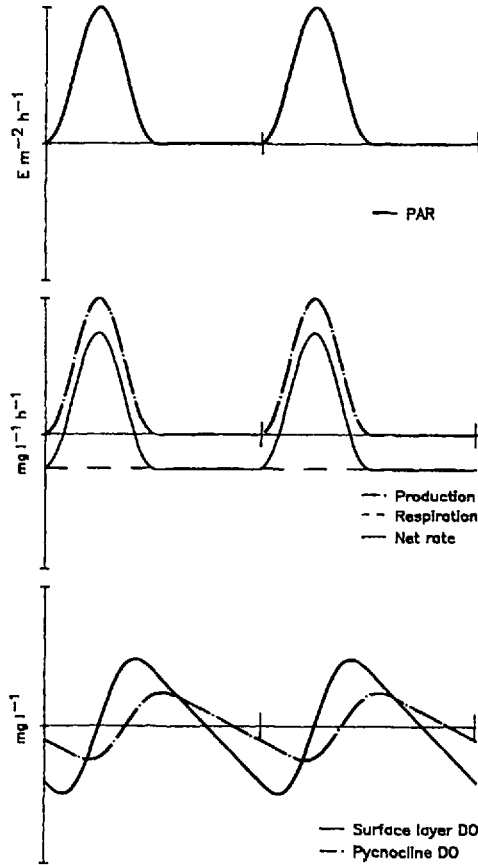


Figure 8. Schematic illustration of DO response to insolation, assuming production is the only source and respiration is the only sink. Ranges of vertical axes are arbitrary, horizontal range represents two days. *Upper panel*—daily cycle of insolation. *Center panel*—Gross and net ranges of production; assuming that algal production (>0) responds directly to insolation, respiration (<0) remains constant, and that the net rate is simply the sum of the two. *Lower panel*—DO response; in the surface layer, this is the integral of the net rate in the center panel; in the pycnocline, the phase lag is increased and the amplitude decreased by downward diffusion.

It should be noted that the hypothetical behavior illustrated in Figure 8 does not necessarily agree with the pattern of insolation-DO phase lags apparent in Figure 5, where deeper stations often did not show greater insolation-DO phase lags than shallower stations. The reason for this discrepancy may have to do with spatial variability in the vertical structure of the water column, such that a shallow station at one site may have been more affected by time dependent downward diffusion than a deeper station at another site. At the only site with DO sensors at multiple depths (CBW 6 m and 9.4 m), the phase lags were not statistically different and the question is unresolved. The possibility of time dependent downward diffusion clearly needs further investigation.

The lack of significant coherence between insolation and DO at mooring CS1 6 m is intriguing. Physically, the site was quite similar to CH 3.7 m; i.e., near the bottom in a shallow, usually well mixed environment subject to occasional intrusions of saline, hypoxic water. The Choptank was slightly more turbid than the mainstem of the Bay during the period of observation, such that the euphotic zone was shallower (KS, unpubl. data) and the sensor may have been more isolated from near-surface diel fluctuations than at an equivalent depth in the mainstem of the Bay. However, data from stations that were at greater depth than CS1 6 m (CBW 9.4 m and CE2 13.1 m) did show significant insolation-DO coherence. Alternatively, it may be that some other factor not examined controlled local production to a greater extent in the Choptank than in the mainstem of the Bay. The low coherence levels between DO and both insolation and salinity at CS1 6 m illustrate the need for time series measurements of other ecologically important variables.

d. High frequency variability. High frequency DO variability that was inversely related to salinity must have been due to internal waves and/or internal mixing. At the very highest frequencies, individual internal waves and mixing events might be observed. Large amplitude (1–2 m), high frequency internal waves (periods of 1 min to 1 h) and internal mixing events have been observed at the pycnocline in the mid-Chesapeake Bay (e.g., Brandt *et al.*, 1985; Dubbel *et al.*; 1985; Sarabun *et al.*, 1985, LS unpublished data). Occurrences often appear to be tidally modulated. At lower frequencies, the effects of mixing might be observed. Cross-spectral analyses of the DO, salinity, and speed records from site CE2 (not presented here) have shown a strong relationship between low salinity, high DO, and high current speed at exactly twice the semi-diurnal tidal frequency. This most probably resulted from enhanced internal mixing across the pycnocline at near peak tidal speeds at mooring CE2.

Advection of horizontal spatial patchiness in DO past the point of observation also would result in high frequency temporal variability in DO. Atkinson *et al.* (1987) examined the patchiness of DO in the Swan River estuary, West Australia. They identified a lower layer horizontal patchiness length scale of about 100–200 m. Though the Swan River is much smaller and more stratified than the Chesapeake Bay, the lower layer environments of the two estuaries are similar. A 200 m patchiness length scale would result in a time scale of 15 min if advected past a fixed mooring at 15 cm s^{-1} , which is characteristic of tidal current speeds in the Chesapeake Bay. Advected patchiness may explain much of the observed high frequency variability of DO that was not related to salinity variability.

Local production and respiration probably were responsible for some of the slower variation in DO, and they may have combined with physical dispersion to produce spatial patchiness (Okubo, 1977), but alone they could not have caused the observed high frequency variability in DO. For example, using the measured benthic oxygen consumption rates of Kemp *et al.* (1987) from site near CBW in August, 1986, and assuming that benthic consumption acts only on the lowest 1 m of the water column,

then a rate of decrease of $1.7 \text{ mg L}^{-1} \text{ d}^{-1} = 1.2 \times 10^{-3} \text{ mg L}^{-1} \text{ min}^{-1}$ might be attributable to benthic consumption alone. This is three orders of magnitude smaller than some of the rates of change in DO observed (Figure 6, about $1 \text{ mg L}^{-1} \text{ min}^{-1}$). Similarly, neither water column respiration rates nor autotrophic oxygen production rates approach the observed rates of change.

Finally, sampling artifacts may have reduced the correlation between salinity and DO at high frequencies. At mooring CBW, for example, salinity was measured on a current meter mooring about 50 m away from the DO mooring, though the sensors were at the same depth. In addition, the DO sensor measurements represent an effective integration over some past time scale determined by diffusion across the membrane, while the salinity measurements are instantaneous. These sampling differences probably had a negligible effect at low frequency and large space scales, but they may have adversely affected the relationship between measured DO and salinity at the highest frequencies and shortest space scales.

5. Conclusions and discussion.

The observations reported here have shown that DO in the mid-Chesapeake Bay during the summer of 1987 was highly variable in time, and that the magnitude and frequency of the temporal variability depended strongly on location. Time scales of variability covered a broad spectrum ranging from the sampling interval (several minutes) to the longest period fluctuations resolved (several days).

The largest variability observed (up to the full range between supersaturation and anoxia) tended to be associated with large amplitude, wind and tide forced lateral internal oscillations of the pycnocline in the mainstem of the Bay, which resulted in advection of saline, hypoxic water from below the pycnocline onto the flanks of the Bay and into the lower reaches of adjoining tributaries. Advective variability at higher frequencies may have been associated with internal waves, with internal mixing, and with spatial patchiness. In general, the magnitude of advective variability decreased as the frequency of the driving physical process increased, but occasional episodes of high frequency variability were quite large. The magnitude of advective variability was greatest near the depth of the pycnocline in the mainstem Bay, while the relative importance of advective variability was largest in the lower layer.

Wind forced lateral oscillations of the pycnocline were driven by reversals in the longitudinal component of the wind. Lateral upwelling and downwelling that accompanied the dominant longitudinal response were manifested by rapid vertical motion of the pycnocline over the flanks of the Bay and corresponding onshore-offshore motion of the benthic front at the intersection of the pycnocline with the Bay bottom. Typical wind forced oscillation cycles occurred every 2–6 d, with vertical pycnocline excursions of about 10 m and corresponding lateral excursions of about 4 km on the western shelf of the Bay near the measurement sites.

Internal tidal oscillations of the pycnocline occurred at both diurnal and semi-diurnal frequencies. These internal tides appeared to result from near resonant forcing

of the lowest mode lateral internal response by surface tides. Maximum wave height was about 5 m, such that the internal tide appeared as an additional 2–3 meter amplitude oscillation superimposed on the wind-forced oscillations of the pycnocline. The existence of a near resonant lateral internal tidal response depends on both the cross-sectional geometry of the Bay and the vertical density structure. As a result, a lateral internal tidal response may be localized and should be modulated as the density structure changes over meteorological and seasonal time scales. Longitudinal components to the internal tide also may be implicated. The lower Choptank River appeared to be isolated from internal tides in the Bay by sills in the entrance channel.

The daily cycle of insolation also caused variability of DO at or near 1 cpd. The influence of insolation was greatest near the surface, as expected. DO responded to insolation with a phase lag of 90° or more. This lag was probably due to the combined effects of the daily cycle of production and loss of DO in the surface layer with time dependent diffusion of DO downwards from the surface layer.

To lowest order, DO variability due to the combined effects of advection and insolation might be predicted by a linear superposition of the various responses found here. For example, our results would predict that the lowest near bottom DO at site CH should occur when a northward wind accompanies a high low tide (high diurnal tide and low semi-diurnal tide) in the morning. This prediction agrees reasonably well with the results of Carter *et al.* (1978) (except for wind direction, see Section 4) and Breitburg (1990), who report on the temporal variability of DO near site CH. However, the large magnitude of some of the advective changes observed is not consistent with simple linear superposition of the various responses. Again using site CH as an example, Breitburg (1990) suggests that the early morning intrusions of subpycnocline waters that produced the most severe declines in DO in 1987 and 1988 probably precluded any simultaneous insolation response because subpycnocline waters in the summertime Bay tend to be aphotic. Thus, some of the relationships found here may indicate sequential, rather than simultaneous responses of DO to insolation and advection. This does not affect the validity of the results for diagnostic purposes, but it does indicate caution in the application of the results for predictive purposes.

The relationship between salinity and DO changes over time because there are a suite of biological and chemical processes that can change DO independently of salinity, and because the boundary conditions for salinity and DO are not the same. The only biological process that could be addressed with the present data set was the insolation response. The influence of other factors was presumably contained within the error of the present analysis, i.e., that part of DO variability which was not explained by variations in salinity and insolation. Advected patchiness, deviations from linearity, and non-stationarity also acted to increase this error. For the most part, the data presented here indicate that processes acting to change the relationship between salinity and DO were of secondary importance relative to advection over the time scales of interest (days to minutes), especially in near-bottom waters. Statistically, DO

variability observed at site CS1 was an exception, though the largest changes in DO were visually correlated to changes in salinity.

Large amplitude, short term advective variability is likely to be an important characteristic of most benthic habitats in the summertime Chesapeake Bay and the lower reaches of adjoining tributary estuaries. In the mid-Bay (from the Potomac River north to the Bay Bridge at Annapolis) approximately 75% of the total bottom area lies between 3–17 m (calculated from data in Cronin and Pritchard, 1975). This is the depth range most affected by wind and tide forced advection of low DO in our data. Other ecologically important variables, such as suspended seston, nutrients, planktonic biomass, dissolved hydrogen sulfide, etc., may behave similarly.

Short term advective variability has different implications for the various organisms that inhabit the mid-Chesapeake Bay. Those organisms that are sessile or are behaviorally tied to some specific location will be affected in much the same way as the moored instrumentation deployed during these studies (e.g., Breitburg, 1988). Planktonic organisms will be unaffected by advective variability to lowest order, though they may be affected by subtler changes associated with the physical processes responsible for advection. Nektonic organisms and migratory zooplankton will only be affected as they are periodically excluded from certain environments by decreased DO levels (e.g., Price *et al.*, 1985; Roman *et al.*, 1988). The timing of periods of large advective variability relative to a particular organism's life cycle may also be important. For example, Breitburg (1988) demonstrated that episodic intrusions of low DO water into shallow western shore habitats could adversely affect the reproductive success of *Gobiosoma boscii*, the naked goby.

It is generally acknowledged that correct interpretation of observations of biological and chemical variability requires consideration of the physical context of measurement. The observations of DO variability reported here illustrate and expand on that point. Consideration of the physical context of measurement must include distinguishing between changes within a given water mass and changes caused by advection of a different water mass past the point of measurement. Short term, anomalous variability is common in sparsely sampled time series resulting from large scale mapping and monitoring efforts (e.g., the USEPA Chesapeake Bay monitoring data base). Our results indicate that advection should be considered a probable cause of such variability in shallow marine environments.

Acknowledgments. We thank B. Albright, K. Braun, M. Bundy, M. Marsh, L. May, C. Moore, J. Ortiz, R. Whaley, and the captains and crews of the RV's *Aquarius*, *Orion*, and *Discovery* for their help in gathering and processing the data reported here. L. Bahner, R. Brinsfield, B. Drake, R. Magnien, M. Roman, and M. Wieland generously provided access to additional data. We thank W. Boicourt and K. Mountford for helpful discussions, and S.Y.-Chao for reviewing an earlier draft of this manuscript. We also thank two anonymous reviewers. This work is the result of research partially funded by the Maryland Department of Natural Resources (contract no. C039-87-025), and partially funded by NOAA Office of Sea Grant, U.S. Department of Commerce, under Grant Nos. R/DO-1B and R/DO-12 to the Maryland Sea Grant College and Grant No. NA85AA-D-SG042 to the Virginia Graduate Marine Science

Consortium and Virginia Sea Grant College Program. The U.S. Environmental Protection Agency also supported this work through Interagency funds transfer DW 13-2707-01. The U.S. Government is authorized to produce and distribute reprints for governmental purposes notwithstanding any copyright notation that may appear hereon. UMSCEES Contribution No. 2140.

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