Wind-Induced Destratification in Chesapeake Bay*

DAVID M. GOODRICH**

National Oceanic and Atmospheric Administration, Stony Brook, NY 11794

WILLIAM C. BOICOURT

Center for Environmental and Estuarine Studies, University of Maryland, Cambridge, MD 21613

PETER HAMILTON

Science Applications International Corporation, Raleigh, NC 27606

DONALD W. PRITCHARD

State University of New York at Stony Brook, Stony Brook, NY 11794

(Manuscript received 20 January 1987, in final form 15 June 1987)

ABSTRACT

Multiyear continuous observations of velocity and salinity in the Chesapeake Bay indicate that wind-induced destratification occurs frequently from early fall through midspring over large areas of the estuary. Storm-driven breakdown of summer stratification was observed to occur near the autumnal equinox in two separate years. Surface cooling plays an important, though secondary, role in the fall destratification by reducing the vertical temperature gradient in the days prior to the mixing event. Large internal velocity shear precedes mixing events, suggesting a mechanism involving the generation of dynamic instability across the pycnocline. Destratification is shown to fundamentally alter the response of the velocity field to subsequent wind forcing; in stratified conditions, response is depth-dependent, while after mixing a depth-independent response is observed.

1. Introduction

Vertical mixing has been shown to have a prominent role in the gravitational circulation of estuaries. Increased exchange between surface and bottom water results in an increase in the potential energy of the system and the generation of horizontal density and pressure gradients of greater magnitude than would be present in a completely stratified system (Pritchard, 1967). In estuaries, the tide is often assumed to be the major source of this mixing energy. Over the last decade, however, there has been an increasing awareness of the importance of wind-driven circulation in estuaries; in many systems, the amplitude of wind-driven flows frequently exceeds that of the lower-frequency gravitational flow. Yet little attention has been devoted to the vertical mixing effect of the wind. In itself, the degree of stratification can be seen as a dynamic balance between buoyancy input, which will tend to increase stratification, and turbulent kinetic energy input, which will tend to break it down. Given the significance of wind-induced motions in estuaries, wind mixing may be an important source of turbulent kinetic energy in this balance.

The vertical density structure will in turn exert a strong influence on many chemical and biological processes operating in estuaries. Phytoplankton populations in particular are strongly regulated by the vertical structure of their environment. Tyler and Seliger (1978) demonstrate that changes in stratification, together with estuarine circulation, play a major role in determining the seasonal distribution of the dinoflagellate *Prorocentrum mariae-lebouriae* in the Chesapeake. The onset of spring stratification is also a major factor in the development of seasonal bottom water anoxia in the middle and upper reaches of the Chesapeake (Officer et al., 1984). A strong pycnocline forms a barrier to the downward transport of oxygen, while reoxygenation follows a marked reduction in stratification in early fall. In addition to promoting reoxygenation of bottom water, destratification will induce changes in the form and vertical distribution of major nutrient species.

This fall destratification is a large-scale phenomenon which has been evident in historical observations for

* Contribution No. 565 of the Marine Sciences Research Center, State University of New York.

** Current address: NOAA Office of Climatic and Atmospheric Research, Rockville, MD 20852.

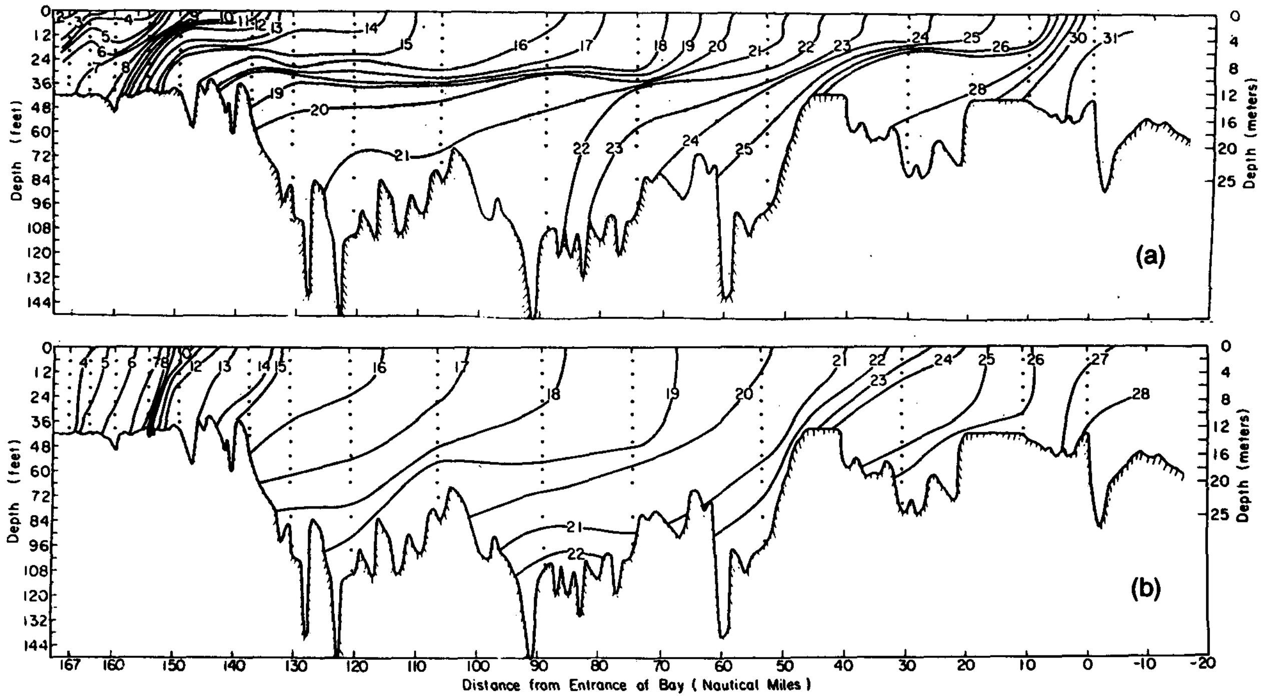

FIG. 1. Longitudinal salinity distribution (ppt) in the Chesapeake Bay (a) 19 September 1968 and (b) 24 October 1968. Reproduced from Seitz (1971).

many years. Figure 1a shows the longitudinal salinity distribution in the Chesapeake in September 1968, indicating a strong halocline throughout the mid-Bay region, which is typical of summer conditions. One month later (Fig. 1b) this stratification has virtually disappeared.

The Chesapeake has been a region where significant advances in our understanding of meteorologically driven circulation have occurred. Using year-long current records in the Potomac River, Elliott (1978) found that the classical two-layer tidal mean flow (seaward-flowing surface and landward-flowing bottom water) was present only 43% of the time. For other periods, the two-layer flow was effectively masked by meteorological effects, generated either locally by wind blowing over the surface of the estuary, or nonlocally by sea level oscillations propagating into the estuary from adjacent water bodies. The local wind effect was examined in more detail by Wang (1979), who observed a depth-dependent response to longitudinal wind in the upper Chesapeake and a depth-independent response in the lower reaches. This flow regime was explained using an analytical model which incorporated wind stress and bottom friction but assumed a homogeneous water column. In general, the presence of a prominent meteorological signal in estuarine current records is widespread, having been found in Narragansett Bay (Weisburg, 1976), San Francisco Bay (Walters, 1982) and Delaware Bay (Wong and Garvine, 1982).

Evidence from studies in fjords suggests that the wind-mixing effect may be important even in summer conditions. In an energy budget for Knight Inlet, Farmer and Freeland (1983) showed mixing energy derived from wind stress in the summer to be one-third of that derived from tidal interaction with bottom topography. This contribution will be dominated by storm events, since wind-energy dissipation on the sea surface varies as the cube of the wind speed. In the mid-Chesapeake, wind mixing should be relatively more important because of the low tidal amplitude (0.3 m) and consequent lower energy available for tidal mixing. Wind mixing would be expected to be less apparent in the lower Chesapeake against the background of stronger tidal mixing there. Indeed, Haas (1977) has reported stratification and destratification on neap–spring periods in the tributary estuaries of the lower Chesapeake.

There is question, therefore, as to the importance of wind mixing to estuarine dynamics and to chemical and biological processes operating in the water column. Because of the event-scale nature of the process, long continuous records are required to insure that records of the events are obtained. In addition, observations from multiple stations are needed to define the spatial scales of the process. From fall 1981 through fall 1983, the National Oceanic and Atmospheric Administration (NOAA) and the Chesapeake Bay Institute (CBI) maintained current meter moorings in the Chesapeake. These data will be used here to analyze the significance and mechanism of the wind mixing process.

2. Observations

During fall 1981, five current meter arrays were deployed in the main stem of the Chesapeake, with each array obtaining surface and bottom measurements of current velocity, temperature and conductivity. Station locations are shown in Fig. 2, with station and instrument depths given in Table 1. The CBI moorings at Fairlee Creek (FC) and Bay Bridge (BB) used Aanderaa RCM-4 current meters; details of the fall CBI experiment are given in Hamilton and Boicourt (1984). Grundy Model 9021 current meters were used at NOAA stations 36, 21 and 2. Salinity transects at Smith Point (Fig. 2) were run in conjunction with these observations in September 1981 using a Grundy 9400 conductivity-temperature-depth (CTD) system. Following the fall 1981 series, NOAA station 36 was continuously maintained for two more years and recovered in December 1983. Where indicated, data have been passed through a Lanczos filter with a 34-hour half-power point in order to remove variance at tidal frequencies and above.

TABLE 1. Current station specifications. Depths are referenced to mean low water.

Station designator	Location	Station depth (m)	Instrument depths (m)
FC	39°16′42″N	7.7	2.4
	76°13′18″W		5.8
BB	38°59′48″N	18.0	2.4
	76°22′08″W		16.8
36	38°18′44″N	17.1	4.6
	76°18′45″W		14.6
21	38°08′06″N	29.7	4.6
	76°13′45″W		12.2
			28.2
2	37°52′45″N	22.5	4.6
	76°09′10″W		12.2
			20.9

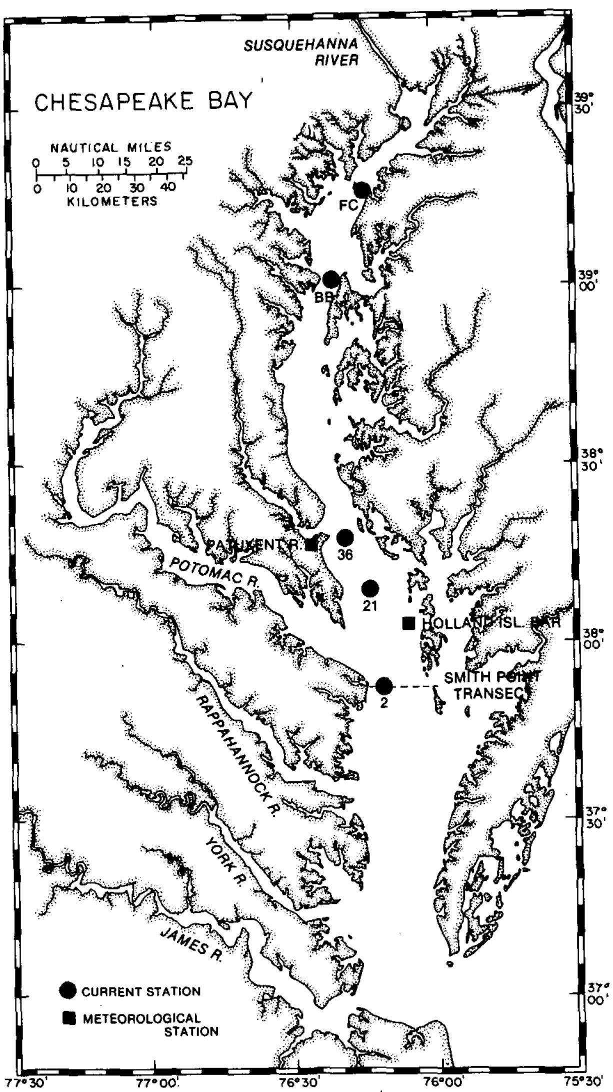

FIG. 2: Station locations.

Wind data for fall 1981 were obtained from a remote meteorological package at Holland Island Bar, a lighthouse 5 km from the nearest land. After this period, data from Patuxent River Naval Air Station, 39 km away, are used. Wind speeds were converted to stress using the quadratic law with a drag coefficient of 1.5 $\times 10^{-3}$ (Pond, 1975). Discharge for the Susquehanna River at Conowingo, Maryland was obtained from the U.S. Geological Survey. This station represents roughly 85% of the total runoff to the Bay north of the Potomac River.

3. Results

Using these wind, current and salinity data, we intend to show that wind-induced, top-to-bottom mixing is a process operating over large areas of the Chesapeake, and one that occurs frequently during fall, winter, and early spring. The transition from the stratified conditions of late spring and summer to these intermittently mixed conditions will be shown to occur suddenly and predictably during the first major storm of early fall. In fact, the sequence of events preceding the initial fall destratification is regular enough that this destratification happened within two days of the same date in two separate years of observation. The long-term data will then be used to make inferences regarding the mechanism of pycnocline breakdown. First, however, we will examine the well-documented case of the initial destratification of fall 1981.

On the afternoon of 22 September 1981, a lateral salinity transect was made at Smith Point in light winds, showing a moderately stratified water column with fresher water to the west, in keeping with rotational effects (Fig. 3a). Five hours after these observations were made, down-Bay (southward) winds increased to

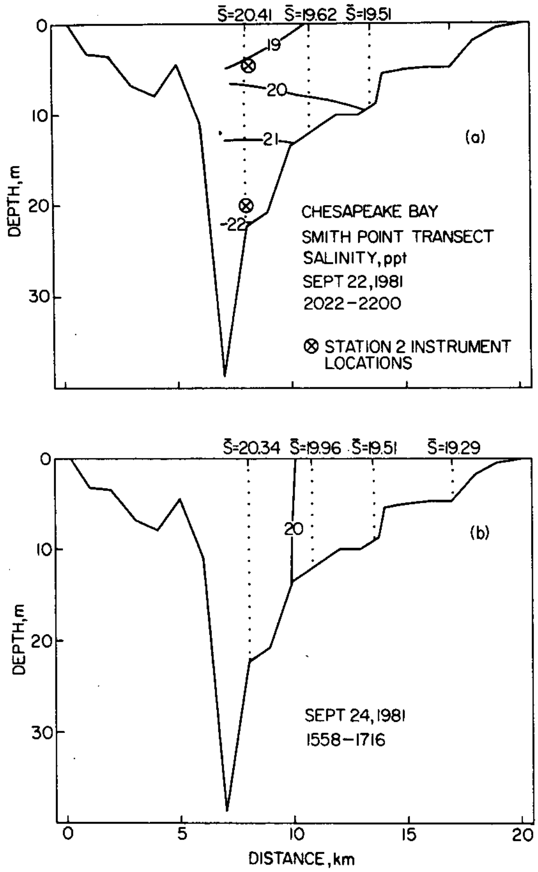

FIG. 3. Lateral salinity distribution off Smith Point looking northward (a) before and (b) after the storm of 23–24 September 1981. Depth-mean salinity indicated above stations.

16 m s^{-1}, and this storm continued for roughly 36 hours. As the winds were abating, the transect was reoccupied on the afternoon of 24 September, with the results shown in Fig. 3b. A vertically homogeneous water column was encountered at each station, extending to at least 22 m in the deep channel. Note that the salinities shown in Fig. 3b are very close to the depth-mean salinities prior to the destratification, an observation consistent with direct vertical mixing rather than longitudinal advection of well-mixed water from elsewhere in the Bay.

Response of the water column before and after the destratification event can be seen by examining the records from the moored instruments at station 2, at the midpoint of the Smith Point transect. Figure 4 shows the salinities and principal axis (essentially north–south) velocities from surface and bottom instruments at this station, together with hourly winds from Holland Island Bar. The semidiurnal tidal current can be clearly seen in the velocity records (Fig. 4b); a mean seaward (positive) surface flow and landward (negative) bottom flow can also be discerned during periods of calm winds prior to the storm of 23–24 September. As the wind increased on 23 September, increased down-Bay surface flow was generated (1), and large shear was present between surface and bottom instruments. During September 24 surface and bottom salinities converged (Fig. 4c), indicating destratification to the depth of the bottom instrument. At this point, surface and bottom velocities became virtually identical (2), despite continued presence of a strong down-Bay wind. This is suggestive of increased frictional coupling in a well-mixed water column, and this interrelationship between the fields of velocity and mass will be examined later using longer time series. Note also that the well-mixed water column was quite transient, with stratification rapidly reestablished as winds decreased.

Records from the five current meter arrays deployed along the Bay axis during fall 1981 indicate that destratified conditions were widespread during this time. Low-pass filtered top-to-bottom salinity difference for the five arrays, deployed over two-thirds of the length of the Bay, are shown in Fig. 5, along with Susquehanna River discharge. Three of the arrays (stations BB, 21 and 36) were located in the main channel. The 7-day cycle evident in the flow records prior to the end of October is a result of weekend shutdown of power generation turbines at the Conowingo Dam; drought conditions were prevalent during this period.

At the beginning of September, stations 36 and 2 show a well-stratified water column. Following the storm of 23–24 September, both stations drop to near zero salinity difference, and stratification at station 36 remained low for a period of one month. In mid-October, all four of the arrays which returned data show destratified conditions, indicating that this was neither

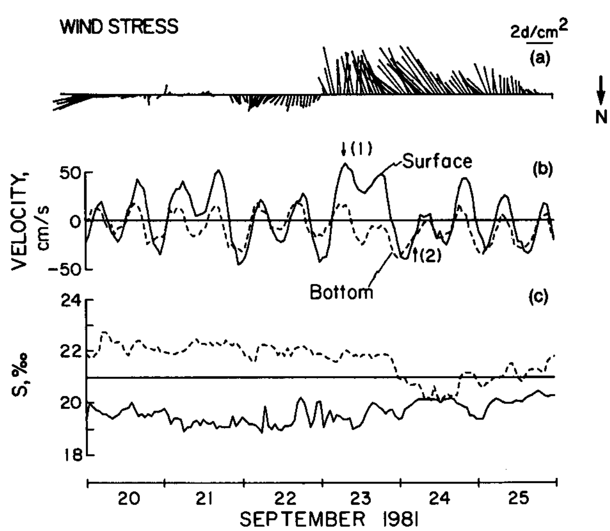

FIG. 4. Wind stress, current and salinity time series off Smith Point, September 1981. All data unfiltered. (a) Wind stress, Holland Island Bar; southward wind to top of page. (b) Station 2 principal axis velocity, surface (solid) and bottom (dashed). (c) Station 2 surface and bottom salinity.

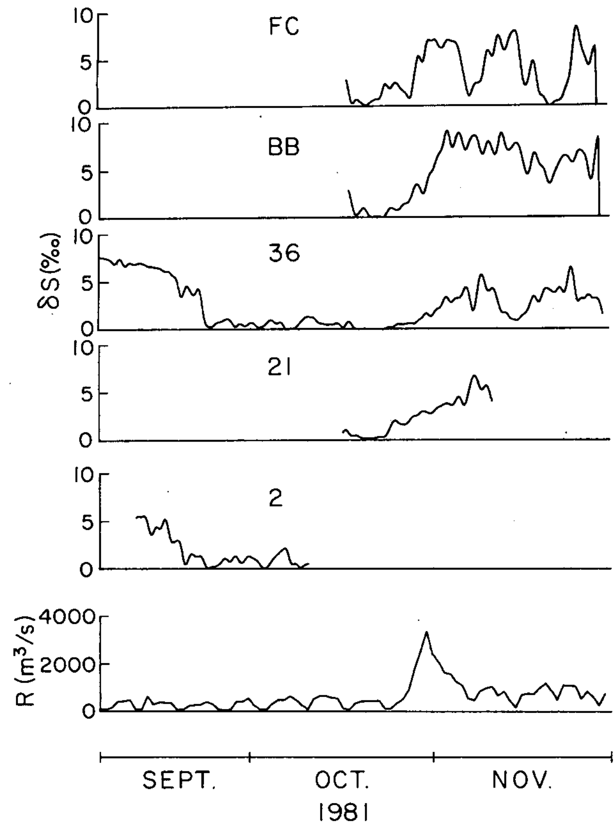

FIG. 5. Filtered top-to-bottom salinity difference at CBI moorings FC and BB and at NOAA moorings 36, 21 and 2, with daily average Susquehanna River discharge; fall 1981. Station depths shown in Table 1.

a local phenomenon nor one confined to shallower waters. This extended even to station 21, which at 28 m was located in one of the deeper reaches of the Bay. Stratification was reestablished in late October, in apparent response to sharply increased discharge from the Susquehanna; stations FC, BB, 21 and 36 show essentially simultaneous restratification.

It is recognized that surface cooling in the Chesapeake in early fall inverts the vertical temperature gradient, which then opposes the salinity-induced density gradient (Seitz, 1971). How important is this inversion in the fall destratification? To examine this process, surface and bottom temperatures at station 36 for September 1981 are plotted in Fig. 6a. Surface temperatures dropped rapidly from 17 to 22 September, falling below bottom values on September 19; following the destratification of 23–24 September, they became virtually identical. To quantify temperature effects on the field of mass at this site, the percent contribution of vertical temperature gradient to vertical density gradient was calculated as follows and plotted in Fig. 6b:

$$P = \frac{\delta\rho(s, t) - \delta\rho(s, \bar{t})}{\delta\rho(s, t)}$$

where

$\delta\rho(s, t)$ observed top-to-bottom density difference
$\delta\rho(s, \bar{t})$ density difference calculated with observed salinities and constant temperature
$\bar{t}$ $(t_s + t_b)/2$.

This figure shows that on 17 September vertical temperature stratification reinforced density stratification by 9%; five days later it was reducing density stratification by 13%. Thus in the week immediately preceding stratification breakdown, temperature effects reduced vertical density gradients by 22%. Following the destratification, density gradients were small, resulting in large fluctuations in P. A virtually identical pattern of surface cooling is shown by temperatures from the same station in the same time period two years later (Fig. 6c). In 1983 the temperature inversion began on 15 September and resulted in a decrease in the vertical density gradient of 20% over a period of about 5 days. Because decreasing air temperatures and increasing wind stress are characteristic of early fall, this temperature inversion is likely to be an annual phenomenon. Note also that the initial destratification in 1983 (22 September) occurred within 2 days of destratification in September 1981, in the vicinity of the autumnal equinox.

The response of the velocity field to subsequent meteorological forcing was markedly affected by these

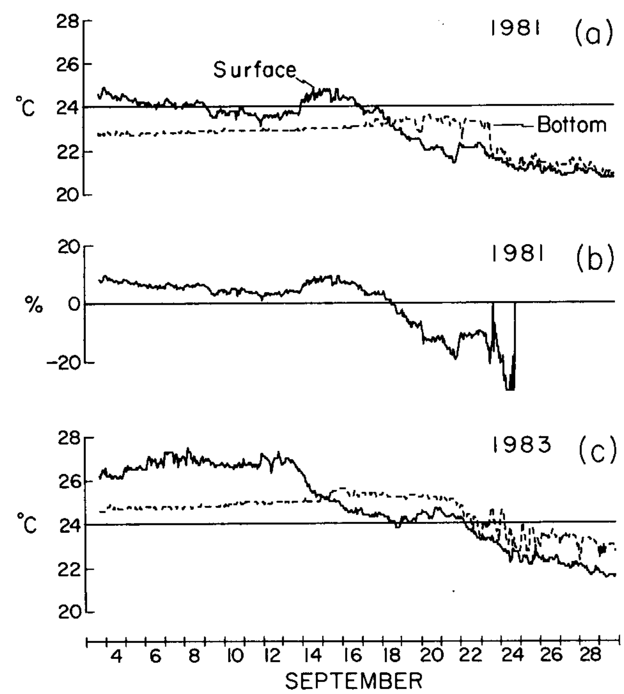

FIG. 6. Fall temperature inversion and effects on density gradient, station 36. (a) Surface and bottom temperature, fall 1981. (b) Percentage of density gradient induced by temperature gradient, fall 1981. (c) Surface and bottom temperature, fall 1983.

changes in the field of mass. Figure 7 shows, for fall 1981, (a) filtered surface and bottom velocities at station 36, (b) surface and bottom salinities, (c) the square of the velocity shear between surface and bottom and, (d) the square of the wind friction velocity (~ wind stress). The subtidal velocity fluctuations at 2–3 day periods were strongly coherent with local winds and can be considered directly wind-driven. The fall destratification is prominent on 23–24 September (spike 3), and this event separates the time series into two very different regimes. In the stratified conditions prior to the 23rd the response was strongly depth-dependent, such that surface current was directly coupled to the wind while bottom current was driven in the opposite direction by the wind-induced slope, after some lag. The three major flow events in this period, 10 (spike 1), 19 (spike 2), and 23–24 (spike 3) September show this response; in addition, each generated large velocity shear and is associated with a deflection in the salinity records. The maximum shear on the 23rd immediately precedes destratification. High shear was also observed at this time further down the Bay at station 2 (Fig. 4).

After the 23rd, the relationship between surface and bottom velocity changed markedly. In the well-mixed water column, the velocities became in phase and response became depth-independent. Cross-spectral analysis confirms this phase shift. For the 17-day period 6–23 September, surface and bottom velocities were 18 hours out of phase at the 2.5-day period, with a coherence squared of 0.78. During the following 17 days, the velocities were just 1.8 hours out of phase at this period, with coherence squared of 0.82. Pycnocline breakdown apparently resulted in far stronger coupling between surface and bottom, and in reduced velocity shear. Repeated strong wind events (Fig. 7d) then served to suppress stratification until the high river discharge of late October.

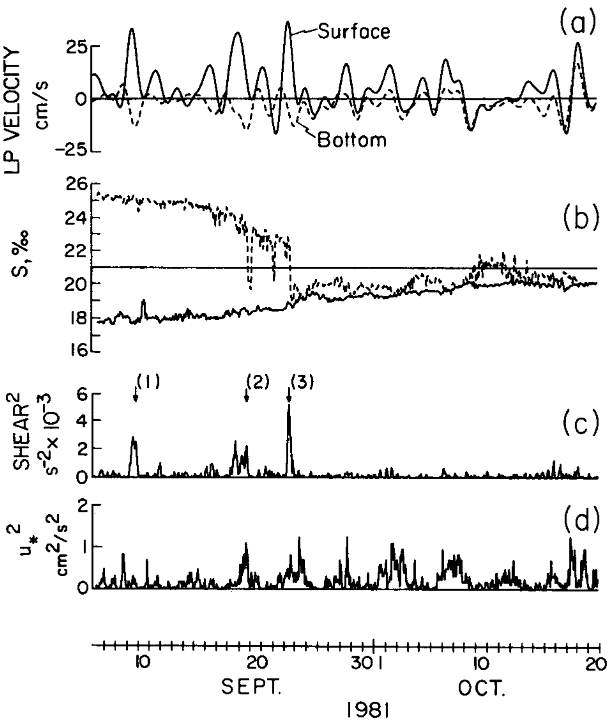

FIG. 7. Current velocity, salinity, velocity shear and wind friction velocity, fall 1981. Labeled events discussed in text. (a) Filtered surface (solid) and bottom (dashed) principal axis velocity, station 36. (b) Unfiltered surface and bottom salinity, station 36. (c) Square of the velocity shear between surface and bottom instruments, station 36. (d) Friction velocity squared (~ wind stress), Patuxent River.

The data shown in Fig. 7 suggests that high internal shear is associated with destratification events and may be involved with the wind mixing process. In studies of mixed-layer dynamics in the open ocean, two potential mechanisms for deepening of the mixed layer have been suggested. In the turbulent erosion scenario (Kato and Phillips, 1969), turbulence generated at the surface propagates downward through the mixed layer to the pycnocline, causing entrainment of denser water from the assumed quiescent underlying water. This process would be related to a friction velocity $u_* = \sqrt{\tau/\rho_0}$, where τ is the wind stress and ρ_0 is a reference density. An alternate mechanism was suggested by Pollard et al. (1973), in which initial mixed-layer development is through turbulent erosion, but the depth of the mixed layer is limited by mean flow shear across the pycnocline δu. This shear-induced dynamic instability then generates turbulence and local mixing at the density interface.

To develop evidence for the relative roles of mean flow shear and direct wind stress in mixing during a period when multiple mixing events occurred, data from spring 1983 is displayed in Fig. 8, using the same parameters as in Fig. 7. Both surface and bottom salinity (Fig. 7b) show a general decline during the spring period in response to increased runoff. This decline was by no means steady; destratification events were occurring often during the early spring period and cannot be considered anomalous. Four of these events were seen: 27 February (1), 2 (2), 13 (3), and 22 (4) March. Each was immediately preceded by peak shear in the current records. While these periods of high shear were correlated with strong winds, the converse was not always true. The wind event of 18 March (5) produced little shear, and no mixing is apparent in the salinity records. Internal shear thus appears better correlated with these mixing events than does friction velocity. In general, there is reason to question the adequacy of moored current meters for measurement of velocity shear in a heavy sea state. However, the collapse of shear with destratification in these observations suggests that there is not a major contamination effect due to waves.

Examination of Figs. 7 and 8 suggests that there exists some threshold value of shear which must be attained before destratification can occur. This threshold would be dependent on the buoyancy initially present

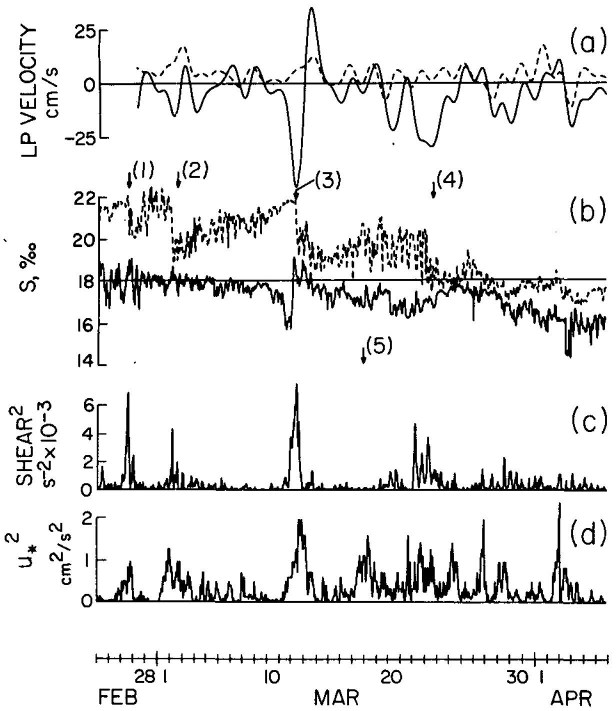

FIG. 8. Current velocity, salinity, velocity shear and wind friction velocity, spring 1983. (a) Filtered surface and bottom velocity, station 36. (b) Unfiltered surface and bottom salinity, station 36. (c) Square of velocity shear, station 36. (d) Friction velocity squared, Patuxent River.

in the system, since more turbulent kinetic energy would be required to mix a more stable water column. A comparison of the stabilizing forces of stratification to the destabilizing forces of velocity shear is the Richardson number

$$\mathrm{Ri} = \frac{g}{\rho_0}\frac{\partial \rho}{\partial z} \Big/ \left(\frac{\partial u}{\partial z}\right)^2$$

where $\partial\rho/\partial z$ and $\partial u/\partial z$ are vertical gradients in density and longitudinal velocity, respectively, g is gravitational acceleration and ρ_0 is a reference density. A bulk Richardson number can be obtained from density difference and velocity shear between the two current meters at station 36. This parameter is shown for two 25-day periods in fall 1981 (Fig. 9b) and fall 1983 (Fig. 9d), together with corresponding plots of salinity difference between top and bottom instruments (Figs. 9a and 9b). Note that the scale for the Richardson number is arbitrarily set at 0–5; large (off-scale) values of Ri are generated in highly stratified periods.

A pattern emerges from the two fall series. As the fall destratification is approached, there are cases in both years when Ri becomes less than 5 for brief periods (associated with wind events), but only when Ri is approximately 0.5 for a sustained period of time does complete destratification occur (arrows 1 and 2). This is suggestive of a critical Ri for large-scale mixing. Laboratory studies (cf. Turner, 1973) indicate a suppression of turbulence for values of the gradient Richardson greater than approximately 0.25. The values shown in Fig. 9 represent a bulk Ri defined over a finite distance, while the concept of a limiting Ri applies to a locally defined value. While the two stability estimates may not be strictly comparable, the similarity of the bulk Ri value at the initial destratification in two different years supports the idea that there exists some critical ratio of buoyancy to velocity shear at which destratification occurs.

This notion is reinforced by the δS–Ri time series for spring 1983 (Fig. 10). The four destratification events shown on Fig. 8 are also indicated here. For each of these events, as in the fall series, a sharp drop in δS is associated with near-zero values of the bulk Richardson number. It should be noted that in both Figs. 9 and 10 that low values of Ri must be sustained for destratification to occur; low values occur intermittently with no apparent effect.

4. Discussion

Continuous salinity records from fall 1981, spring 1983, and fall 1983 indicate that top-to-bottom wind-induced mixing occurs often during these seasons, and records from multiple arrays show that this process extends over most, if not all, of the Bay. This is some-

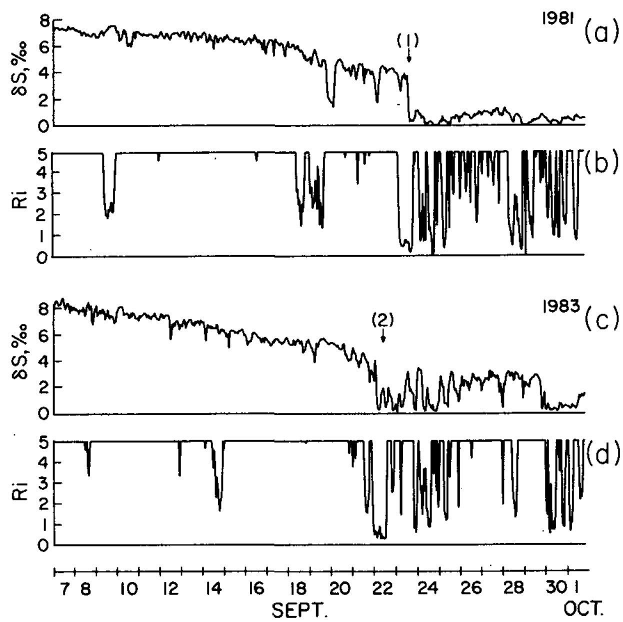

FIG. 9. Stratification and Richardson number, fall 1981 and fall 1983, station 36. (a) Surface-to-bottom salinity difference, fall 1981. (b) Richardson number, fall 1981. (c) Surface-to-bottom salinity difference, fall 1983. (d) Richardson number, fall 1983.

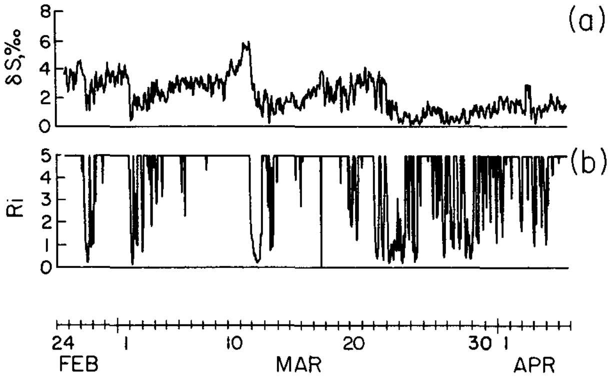

FIG. 10. Stratification and Richardson number, spring 1983, station 36. (a) Surface-to-bottom salinity difference. (b) Richardson number.

what surprising, since historical section data very rarely show completely destratified conditions. An explanation lies in the sampling bias of hydrographic observations taken from ships. Periods of intense wind mixing are likely to coincide with weather conditions which would preclude operation of smaller research vessels. Gravitationally driven restratification is a rapid process following abatement of the wind (Fig. 4), making shipboard observation of deep, well-mixed conditions following a storm difficult.

A special case is the initial fall destratification, which is of particular interest since it results in the reaeration of bottom water following the summer hypoxia. In both fall 1981 and fall 1983, this event occurred within two days of the autumnal equinox. While the close timing between the two years may be somewhat fortuitous, it does indicate that external processes, i.e. surface cooling and wind mixing from fall storms, can be expected to produce destratification sometime around the equinox. Of these two processes, surface cooling appears to play a secondary role. The temperature inversion which occurs in late September reduces the vertical density gradient by only 20%–25%, in agreement with the observations of Seitz (1971). This decrease in buoyancy of the surface water reduces the amount of mixing energy required for destratification, and in both 1981 and 1983, a fall storm mixed the water column within a week of the temperature inversion.

In general, the time-dependent stratification seen in the mid-Chesapeake is the result of a dynamic balance between buoyancy flux, induced primarily by the gravitational circulation, and the mixing energy imparted by the wind. The role of gravitational circulation is well illustrated by the fall 1981 series (Fig. 5). In the drought conditions of September–October, the relatively small amount of buoyancy present in the estuary allowed the system to remain well-mixed for a period of one month. The increased fresh water flow in late October generated intensified gravitational circulation, which led to rapid restratification over a large area of the Bay. The response is similar to that observed following Hurricane Agnes in 1972, when record discharge was followed by strong intrusion of saline bottom water (Schubel et al., 1976). In terms of mixing energy, wind stress appears to be the major time-dependent source in the mid-Chesapeake and may be the dominant absolute source.

A scenario for wind-induced mixing is suggested by the prominent role of internal shear, which precedes mid-Bay mixing events in virtually every case. A longitudinal wind with particular characteristics of amplitude and duration drives surface water downwind and sets up a sea surface slope. The barotropic pressure gradient thus generated drives water upwind, inducing shear across the pycnocline. If this shear reaches a critical level with respect to buoyancy across the interface, dynamic instability is generated and large-scale vertical mixing occurs. This suggests a mechanism involving internal waves, and breaking internal waves have indeed been observed acoustically in this region of the Chesapeake (Sarabun et al., 1984). Resolution of these processes is not possible from a two-instrument mooring. Regardless of the exact mechanism of pycnocline breakdown, the importance of internal shear in the mixing process is supported by other studies, notably that of Price et al. (1978). In observations of density and velocity structure during storms in the Gulf of Mexico, mixed-layer deepening was found to cease abruptly as shear across the pycnocline decreased, despite increasing wind stress. The implication for modeling from these observations is that vertical mixing should be related to internal shear, ideally through a dynamic stability (Ri) dependence in the mixing parameterization.

The reduction in shear following mixing implies stronger frictional coupling between surface and bottom water in a well-mixed water column, and this accounts for the shift from depth-dependent to depth-independent response in fall 1981. This stability-dependent response is not unfamiliar; for example Chuang et al. (1979) found baroclinic or barotropic response to be strongly influenced by stratification at a current station in the Middle Atlantic Bight. Given the stronger vertical density gradients in an estuary, more dramatic changes in response following mixing are not unlikely. Models of estuarine response to local wind forcing to date have assumed a homogeneous water column (e.g., Wang, 1979; Elliott, 1982). From the results of this study, however, it is clear that variations in the field of mass will strongly affect the response of the velocity field.

One would not expect wind-induced destratification to be unique to the Chesapeake, although there are topographic influences in this estuary which probably make the effect more prominent. The long fetch present in the Chesapeake allows the development of substantial wind-driven internal shear, whereas in a system with large bends and other highly variable topography, much of this energy would be channeled into generation of secondary flows. However, significant wind mixing should be discernable in other large estuaries such as Delaware Bay or San Francisco Bay, although the greater magnitude of tidal mixing in these systems may mask this contribution from the wind.

In the Chesapeake or elsewhere, wind mixing is a mechanism for the vertical redistribution of chemical species and plankton. Episodic mixing of nutrient-rich bottom water into the euphotic zone has been associated with phytoplankton blooms in a number of estuaries, including the York River (Haas et al., 1980) and the St. Lawrence estuary (Sinclair, 1978). A similar response could be postulated for the main stem of the Chesapeake following a mixing event, assuming adequate light penetration and temperature. In addition, reoxygenation of bottom water following summer oxygen depletion probably occurs during the late September destratification. Similarly, during development of stable stratification in spring, the timing of the last wind-induced water column aeration may have a major influence on the severity of oxygen depletion during the following summer.

In summary, wind-induced destratification events have been shown to occur often and over large areas of the Chesapeake from early fall through spring. These mixing events will alter the response of the velocity field to subsequent wind forcing and can be expected to induce large vertical fluxes of nutrients, dissolved oxygen and other water column constituents. Clearly, the assumption of steady state conditions in the analysis of data taken during this intermittently mixed season will be misleading. Rather, a conceptual model of event-scale mixing followed by gravitationally driven restratification should be used in interpreting discrete observations of spatial property distributions in the Chesapeake. It should be recognized that these distributions will be strongly influenced by the amount of time since the last mixing event.

Acknowledgments. NOAA data was obtained by the NOAA Ship *Ferrel* as part of the Chesapeake Bay Circulation Survey. One of us (D.G.) was supported by the Office of NOAA Corps during the preparation of this paper. P.H. and W.B. were supported by the Maryland Power Plant Siting Program under Contract P81-81-04.

REFERENCES

Chuang, W-S., D-P. Wang and W. Boicourt, 1979: Low-frequency current variability on the southern Mid-Atlantic Bight. *J. Phys. Oceanogr.*, **9**(6), 1144–1154.

Elliott, A., 1978: Observations of meteorologically induced circulation in the Potomac estuary. *Estuarine Coastal Mar. Sci.*, **6**, 285–299.

——, 1982: Wind-driven flow in a shallow estuary. *Oceanol. Acta.*, **5**(1), 7–10.

Farmer, D., and H. Freeland, 1983: The physical oceanography of fjords. *Progress in Oceanography*, Vol. 12, Pergamon 147–220.

Haas, L., 1977: The effect of the spring-neap tidal cycle on the vertical salinity structure of the James, York and Rappahannock Rivers, Virginia, U.S.A. *Estuarine Coastal Mar. Sci.*, **5**(4), 485–496.

——, S. Hastings and K. Webb, 1980: Phytoplankton response to a stratification-mixing cycle in the York River estuary during late summer. *Estuaries and Nutrients*, B. Neilson and L. Cronin, Eds., Humana Press, 619–636.

Hamilton, P., and W. Boicourt, 1984: *Long-term salinity, temperature and current measurements in upper Chesapeake Bay.* Maryland Dept. of Natural Resources Power Plant Siting Program Tech. Rep. PPRP-87, 106 p.

Kato, H., and O. Phillips, 1969: On the penetration of a turbulent layer into a stratified fluid. *J. Fluid Mech.*, **37**, 643–655.

Officer, C., R. Biggs, J. Taft, L. Cronin, M. Tyler and W. Boynton, 1984: Chesapeake Bay anoxia: origin, development and significance. *Science*, **223**, 22–27.

Pollard, R., P. Rhines and R. Thompson, 1973: The deepening of the wind-mixed layer. *Geophys. Fluid Dyn.*, **3**, 381–404.

Pond, S., 1975: The exchange of momentum, heat and moisture at the ocean–atmosphere interface. *Numerical Models of Ocean Circulation*, Natl. Acad. Sci., Washington, DC, 26–38.

Price, J., C. Mooers and J. Van Leer, 1978: Observation and simulation of storm-induced mixed layer deepening. *J. Phys. Oceanogr.*, **8**, 582–599.

Pritchard, D., 1967: Observations of circulation in coastal plain estuaries, *Estuaries*, G. Lauff, Ed., Amer. Assoc. Adv. Sci., 37–44.

Sarabun, C., A. Brandt and M. Tyler, 1984: Breaking internal waves in the Chesapeake Bay (abstract). *Eos Trans. AGU*, **65**(45), p. 907.

Schubel, J., H. Carter and W. Cronin, 1976: Effects of Agnes on the distribution of salinity along the main axis of the Bay and in contiguous shelf waters. In Chesapeake Research Consortium, *The Effects of Tropical Storm Agnes on the Chesapeake Bay Estuarine System*, The Johns Hopkins University Press, 33–65.

Seitz, R., 1971: Temperature and salinity distributions in vertical sections along the longitudinal axis and across the entrance of the Chesapeake Bay. Chesapeake Bay Institute Graphical Summary No. 5, The Johns Hopkins University, 99 p.

Sinclair, M., 1978: Summer phytoplankton variability in the lower St. Lawrence estuary. *J. Fish. Res. Bd. Can.*, **35**, 1171–1185.

Turner, J., 1973: *Buoyancy Effects in Fluids.* Cambridge University Press, 367 pp.

Tyler, M., and H. Seliger, 1978: Annual subsurface transport of a red tide dinoflagellate to its bloom area: water circulation patterns and organism distributions in the Chesapeake Bay. *Limnol. Oceanogr.*, **5**(2), 227–246.

Walters, R., 1982: Low-frequency variations in sea level and currents in South San Francisco Bay. *J. Phys. Oceanogr.*, **12**, 658–668.

Wang, D-P., 1979: Wind-driven circulation in the Chesapeake Bay, winter 1975. *J. Phys. Oceanogr.*, **9**, 564–572.

Weisberg, R., 1976: The nontidal flow in the Providence River of Narragansett Bay: a stochastic approach to estuarine circulation. *J. Phys. Oceanogr.*, **8**, 225–232.

Wong, K-C., and R. Garvine, 1984: Observations of subtidal, wind-induced variability in the Delaware estuary. *J. Geophys. Res.*, **89**(C6), 10 589–10 597.