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Mixing processes on the Atlantic continental shelf, Cape Cod to Cape Hatteras

Hugo B. Fischer
Department of Civil Engineering, University of California, Berkeley 94720

Abstract

Data from moored current meters and airborne and satellite observations are used in conjunction with analyses of shear flow dispersion to estimate the dispersion of dissolved substances in the Middle Atlantic Bight during unstratified periods. The long-shelf dispersion coefficient is estimated to be of the order of 10^4–10^5 $cm^2 \cdot s^{-1}$, and the cross-shelf dispersion coefficient of the order of 3×10^5 $cm^2 \cdot s^{-1}$ at midshelf. The cross-shelf estimate is a factor of 10 less than previous ones, which were based on the assumption that the cross-shelf flux of freshwater is equal to the inflow from tributary rivers within the bight itself.

The Middle Atlantic Bight is about 1,000 km long and generally terminates at the 200-m depth contour, about 100 km offshore. The bight receives the drainage of a substantial fraction of urban America. Among other uses sewage from New York City is barged into the bight, and power stations are proposed to be sited on islands built in it (*see* Gross 1976). Obviously the capacity of the waters of the bight to receive, transport, and remove pollutants is of interest.

The appropriate analysis for dispersion of a substance depends on its lifetime, because substances with different growth, decay, or reaction times respond primarily to different dispersion mechanisms. Short-lived substances, having lifetimes up to 10 days, include such parameters as coliform bacteria, nitrites, nitrates, and BOD; examples of important long-lived substances are phosphates and carbon. The characteristic times of the physical processes are: 0.5–1 day, tidal and inertial periods; 3–10 days, approximate period of events related to the passage of large storms; 1–6 months, seasonal variations in temperature and salinity structure and runoff from rivers; >1 year, year-to-year variation. The study of dispersion can be organized according to the relationships between characteristic times. If the lifetime for the constituent is much less than the characteristic time for the physical process, the physics may be regarded as essentially constant. For instance, the dispersion of coliform bacteria (lifetime ≈ 12 h) in a 3-day storm proceeds much as though the storm had been going on forever, and in a calm period as though the calm had been going on forever. On the other hand, if the lifetime for the constituent is much greater than the characteristic time of the physical process, averaging of the physical variation is permissible. For instance, to analyze the dispersion of BOD (lifetime ≈ 10 days) we can average the effects of the tidal flow over the tidal period, unless dispersion very near the source is of interest. Cases that fall between the extremes are usually the most difficult to analyze; for instance, dispersion of BOD depends on the time history of a passing storm, and dispersion of carbon is affected by seasonal transitions.

It is important to note that knowledge of the dispersion of a short-lived tracer does not necessarily imply adequate information about a long-lived substance, and vice versa. Radon-222 is a natural tracer occurring along the continental shelf with a decay time of about 4 days (Biscaye et al. 1978); the observed distribution of radon from a localized source is a good indicator of the expected distribution of any other substance with a similar lifetime, such as BOD, but it provides little information about the expected dispersion of a long-lived substance such as carbon.

Here I discuss the physical transport and diffusive processes occurring in the Middle Atlantic Bight. My description is limited because the data are limited; my quantitative results are limited to the region inshore of the shelf break and offshore of the region described by Csanady (1975) as the coastal boundary layer, and to periods of the year when the water column is unstratified (approximately November to mid-April). It seems as yet impossible to give a quantitative analysis for periods of stratification, although a partial qualitative description is possible from data now available.

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Data review

Before 1970 what was known of circulation on the continental shelf was inferred primarily from measurements of temperature, salinity, and travel of surface drift bottles and seabed drifters. Sections given by Ketchum and Corwin (1964), south of Montauk Point, and by Beardsley and Flagg (1976), south of Martha's Vineyard, show approximately constant values of density of 1.026 $g \cdot cm^{-3}$ in winter all across the shelf; in summer isopycnal lines are nearly horizontal, and density ranges from 1.023 $g \cdot cm^{-3}$ at the surface to 1.026 at the bottom. Bumpus (1973) used the results of drift bottle studies to construct monthly charts of surface currents. He found a two-cell system, the westernmost of which develops between Hudson Canyon and Cape Hatteras and (p. 136) "receives the outflow from the major rivers emptying into the middle Atlantic bight, including the Connecticut, and continues with a southerly flow toward Cape Hatteras occasionally reaching 15 nautical miles/day, but for the most part restricted to 10 nautical miles/day or less."

Bumpus (1973, p. 150) noted that drift bottles and seabed drifters provide only "a birth notice and an obituary with no biography," and called for a program of continuous current measurements. Such a program has been undertaken, and results are now available for a few periods in a few locations. Figure 1 shows the locations and dates of observation of available data. Most of these data are summarized by Beardsley et al. (1976). The mean velocity vectors at offshore stations are generally alongshore in the direction from Cape Cod to Cape Hatteras, with magnitudes in the range of 3–10 $cm \cdot s^{-1}$. Currents generally increase in magnitude offshore and decrease and veer toward shore with closeness to the bottom. Beardsley et al. noted that at all sites except one near the mouth of Chesapeake Bay there is a net southwestward transport. The current is highly variable depending on the passage of storms, as shown by progressive vector diagrams for the southern half of the bight (Boicourt and Hacker 1976) and for stations off Martha's Vineyard (Mass. Inst. Technol. Data Rep. 76-1). Boicourt and Hacker found velocities >40 $cm \cdot s^{-1}$ for periods as long as 2½ days, leading to longshore travel during a storm of the order of 100 km. Between storms the random wanderings of the progressive vectors often lead to little mean motion. Hence the "mean currents" plotted by Beardsley et al. (1976) are the long term average of widely separated but intense short term events.

Infrared photographs of the eastern seaboard have been taken twice daily since 1973 by NOAA satellites, and the U.S. Coast Guard Airborne Radiation Thermometer Program makes monthly flights. A typical Coast Guard flight makes about nine tracks over the Middle Atlantic Bight from the coast as far seaward as the edge of the Gulf Stream. The results are used to prepare contour maps of surface temperature, at a 1°C contour

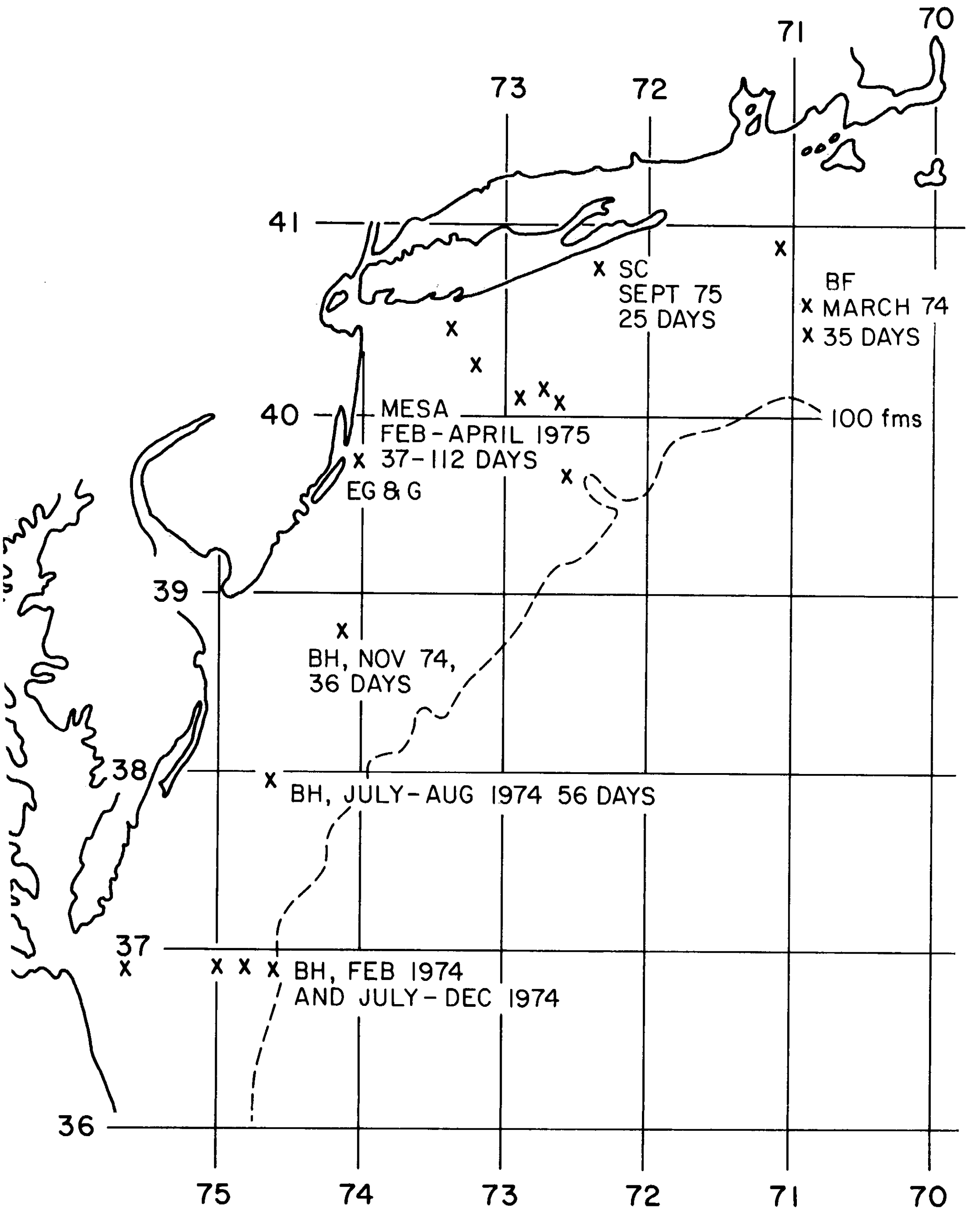

Fig. 1. Locations of moored current meters, giving month of deployment and length of record if known. BH—Boicourt and Hacker 1976; EG&G—observations by EG&G Corp. (*cited in* Beardsley et al. 1976); SC—Scott and Csanady 1976; BF—Beardsley and Flagg 1976.

interval. The NOAA photographs are not contoured and are difficult to interpret in as much detail as the Coast Guard maps, but the photographs are available much more frequently and are synoptic, whereas the Coast Guard must infer the shape of the contours from measurements along the tracks.

The satellite photographs make clear that there are occasional large movements of shelf water offshore across the shelf break. Figure 2 shows the best example I could find (21 November 1975) in the period April 1974 through March 1977. Figure 3 shows the corresponding Coast Guard contour map for 18–20 November 1975. The photograph clearly shows two gyres of shelf water, one south of Cape Cod and one off the mouths of the Delaware Estuary and Chesapeake Bay, each having a diameter of about 1° of latitude. The landward edge of each gyre is at the shelf break. Photographs taken at other dates often show a gyre south of Cape Cod, approximately at the location of and resembling the northern one of the two shown. It is much less common to find gyres in the location of the southern one shown.

The remote sensing data seem to give instant understanding of an important mixing process, but what can we learn from them quantitatively? So far not much. A great deal could be learned if we had shipboard measurements of velocities and subsurface properties within the gyres, but for the moment we can only make some gross estimates based in part on speculation. In the case of the gyres shown in Fig. 2 (but only in this case; no similar sequence was found in the period April 1974–March 1977), it was possible to trace the growth of each gyre through successive photographs from 5 through 21 November. During the period 6–16 November the surface area of shelf water in the southern gyre grew by about 1.3×10^{10} m^2; during the period 6–21 November the northern gyre grew about 6.4×10^{9} m^2. To convert these growth rates to fluxes of shelf water we need to know the depth to which the tongues of shelf water extend. No such

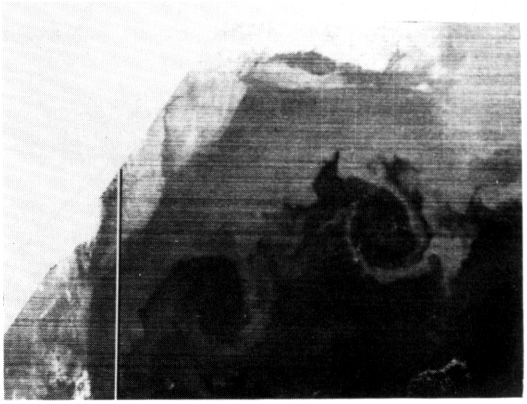

Fig. 2. NASA photograph of continental shelf on 21 November 1975. Cape Cod is at upper right, Cape Hatteras at lower left. Lighter color indicates colder water (*see isotherms shown in Fig. 3*).

measurements are available for November 1975, but we can make a guess based on the results of several previous studies.

The tongues of shelf water shown in Fig. 2 (and in many photographs not shown) appear to result from the impingement of Gulf Stream rings on the shelf break, as described by Saunders (1971) and Morgan and Bishop (1977). Saunders described observations of a single ring, or anticyclonic eddy, which was pinched off from a meander of the Gulf Stream in November 1969. Formation of the eddy from the Gulf Stream was followed during September and October. On 17 November the eddy had become elongated in the east–west direction and had dimensions of about 100 × 200 km at the surface and 60 × 120 km at the 200-m depth. Surface temperature measurements on 17 November show a tongue of shelf water, having a temperature <12°C, lying along 69°W long between 38°30′ and 40°N lat. This is the exact location where tongues of shelf water are seen most frequently in recent satellite photographs and is the location between 6 and 16 November of the tongue which bent around into the northern of the two gyres shown in Fig. 2 on 21 November 1975. Saunders gave a temperature cross section on 17 November 1969, along 39°10′N lat between 68° and 72°W long.

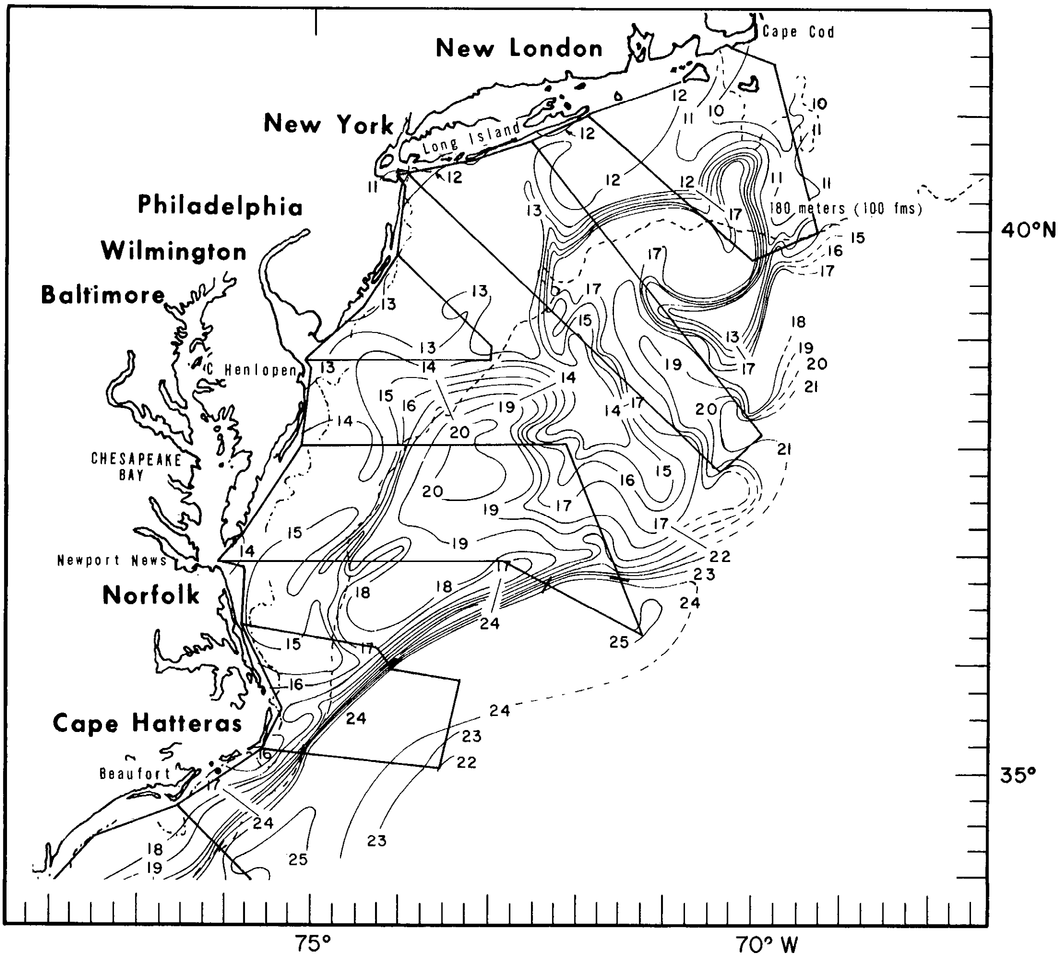

Fig. 3. Surface isotherms in °C measured by U.S. Coast Guard Airborne Radiation Thermometer Program on 18, 19, and 20 November 1975. Solid lines show trace of aircraft; isotherms are interpolated from measurements along tracks.

Between 68°30′ and 69°W a band of water having a temperature between 12° and 10°C was found between the surface and a depth of about 140 m. (fig. 5: Saunders 1971). Saunders described this band as shelf water on the eastern edge of the eddy where it had been displaced at least 150 km to the south (i.e. south of the shelf break, which runs east–west in this vicinity.) It is not clear how much of the water to the 140-m depth is shelf water, since salinities were not measured concurrently with temperature, but much of it is probably from the shelf.

A similar tongue, located near the position of the southernmost of the tongues shown in Fig. 2, is described by Morgan and Bishop (1977). This tongue also lies along the northeast side of a Gulf Stream ring. The observations were made during a cruise of the USCGC *Evergreen* in August 1974; exact dates when the data were obtained are not given. Calculated and observed velocities within the ring were in the range of 30–40 $cm \cdot s^{-1}$ and Morgan and Bishop stated that the shelf water extended at least 50 nm to seaward of the 100-fathom curve. Synoptic salinity plots indicate that the tongue was about 10 km wide; salinity-temperature profiles at two stations within the tongue indicate that it was about 35 m deep but that most of the salinity variation occurred in the uppermost 10 m. Morgan and Bishop computed a flux of shelf water by the dynamic method (i.e. by postulating the hydrostatic and geostrophic balances) of about $10^4\ m^3 \cdot s^{-1}$.

Wright (1976) examined some 19,000 bathythermograms and 1,600 oceanographic stations in the region 39°–41°N lat, 69°–72°W long looking for evidence of changes in the character and position of the shelf water/slope water boundary. He found what he called (p. 1) "detached parcels of shelf water" in the slope water at all times of the year. It seems clear from his methodology and definition of shelf water that many of the "detached parcels" actually represent observations taken in one of the tongues of shelf water now apparent in the satellite pictures. Wright identified two types of parcel: those identified by temperature at the sea surface (type A) and those identified only by subsurface observations (type B). He showed convincingly that the type B parcels were simply type A parcels with surface temperature signature erased by seasonal warming. Wright found that in type A the temperature minimum typically extended to 50–80 m with little change, but in type B the minimum was around 50 m.

These reports suggest that the tongues of shelf water shown in Fig. 2 are between 30 and 140 m thick. If the smaller figure is used as a conservative minimum, a flux of shelf water of $4.5 \times 10^5\ m^3 \cdot s^{-1}$ is computed for the southern gyre and $1.5 \times 10^5\ m^3 \cdot s^{-1}$ for the northern one. These values are substantial compared to the estimates of longshore and offshore flux given below. However, the fluxes estimated in this report may be a maximum for the 3-year period of record. Lesser tonguelike flows do seem common. It is hard to estimate their frequency from the satellite pictures, many of which are missing or show only clouds. The Coast Guard maps suggest that tongues of some magnitude are usually present. Of 26 maps prepared between February 1975 and April 1977, 17 show substantial regions where the contours of constant temperature are perpendicular to the shelf break (indicating a region of offshore flow), two do not show such regions, three are inconclusive, and in four observations were not taken. Subject to confirmation by further study, the tongue observed by Saunders appears typical in size of an event occurring frequently along the northern section of the bight, the tongue observed by Morgan and Bishop was smaller and occurred in a place where large tongues are less frequent, and the tongues shown in Fig. 2 are the grandfathers of the species and may not occur more frequently than once in 2 or 3 years.

Prediction of mixing coefficients

Background—Methods suitable for predicting the values of mixing coefficients on the continental shelf are given in Fischer et al. (1979). The needed results are summarized as follows.

Vertical mixing in turbulent, unstratified, nonrotating, open channel flow can be described by a Fickian mixing coefficient, or eddy diffusivity, whose depth-averaged value is

$$E = 0.067hu^* \qquad (1)$$

in which h is the depth of flow and u^* is the shear velocity ($\sqrt{\tau/\rho}$ where τ is the shear stress and ρ is the fluid density) at the channel bottom. In rotating flows the depth of stress gradients is limited by the formation of Ekman layers of approximate thickness $0.1u^*/f$, where f is the Coriolis parameter and u^* is the shear stress at the top of the Ekman layer in the case of a wind-driven flow. For flows deeper than the Eckman layer, Csanady (1976) suggested replacing the depth by the thickness of the Ekman layer to give, with some rounding off of the coefficient,

$$E = u^{*2}/200f. \qquad (2)$$

If a cloud of tracer substance is deposited at the bottom of a flow in which the vertical mixing coefficient is everywhere E, the time required for diffusion to bring the surface concentration to within 10%

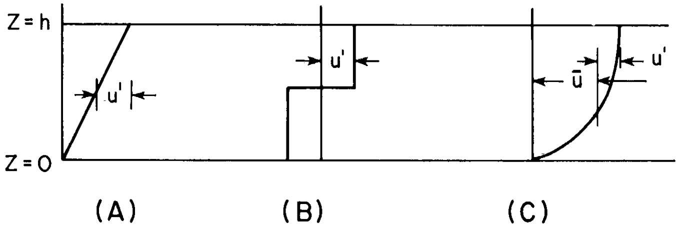

Fig. 4. Velocity profiles. A—Linear profile, $k = 1/30$; B—step profile, $k = 1/12$; C—logarithmic profile $u = \bar{u} + (u^*/\kappa)[1 + \ln(z/h)]$, $k = 1/15$.

of the bottom concentration is given approximately by

$$T_c = 0.4h^2/E. \tag{3}$$

Horizontal mixing is caused primarily by the interaction of vertical mixing with the vertical variation of the horizontal velocity, a process known as shear flow dispersion and first analyzed by Taylor (1954) with respect to dispersion of matter in turbulent flow through a pipe. A general result for two-dimensional shear flows (like those illustrated in Fig. 4) is

$$D = ku'^2h^2/E \tag{4}$$

in which k is a coefficient whose magnitude depends on the details of the velocity profile and the variation of the mixing coefficient, and u' is a characteristic value of the deviation of the horizontal velocity from its vertical mean. Values of

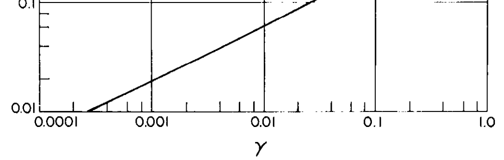

Fig. 5. Effect of period of oscillation on longitudinal-dispersion coefficient caused by shear effect. γ is ratio of dispersion coefficient in a flow oscillating with period T to dispersion coefficient in same flow as $T \to \infty$.

k for the velocity profiles shown in Fig. 4 can be computed as follows: linear profile, $k = 1/30$; step profile, $k = 1/12$; logarithmic profile $u = \bar{u} + (u^*/\kappa)[1 + \ln(z/h)]$, $k = 1/15$. The first two results (Fig. 4A, B) are based on a constant value of E throughout the water column; the third (Fig. 4C) is Elder's (1959) result, which uses a parabolic distribution of the mixing coefficient. Usually little error results from assuming a constant value.

In an oscillatory shear flow, like that shown in Fig. 4C but with $u' = (u^*/\kappa)[1 + \ln(z/h)]\sin(2\pi t/T)$, the horizontal mixing coefficient is reduced by a factor depending on the ratio $T{:}T_c$ where T is the period of oscillation (*see* Fischer et al. 1979). Figure 5 shows the reduction factor.

If the velocity vector has x and y components which vary independently with depth but not with time, i.e. $\vec{u} = u(z) + v(z)$, the same method of analysis that led to Eq. 4 can be used to obtain a dispersion tensor,

$$D_{ij} = ku'_iu'_jh^2/E \tag{5}$$

where u'_i and u'_j are characteristic values of the velocity deviation in the i and j directions. The values of k_{xx} and k_{yy} are determined only by the components of the velocity profiles in the x and y direction, in the same way as k in Eq. 4 is determined in unidirectional flow; the value of k_{xy} depends on the interaction of the velocity profiles and can be negative. To my knowledge the case of an oscillatory shear flow with depth-varying direction of the velocity vector has not been analyzed.

Use of Taylor's method is only valid if

the flow is uniform and steady (or quasi-steady in the case of a pulsating or oscillating flow) for a period longer than the time required for vertical mixing. Csanady (1976) estimated $E = 100\ \mathrm{cm^2 \cdot s^{-1}}$, based on an average magnitude of the surface wind stress; Hunter et al. (1977) obtained the same value by fitting an Ekman model to velocities observed at the 38-m depth. This result and Eq. 3 give mixing times of about 1 day at the 50-m depth and 4 days at the 100-m depth during periods of no stratification. Except near the Hudson Canyon the topography is reasonably uniform. Therefore the continental shelf, away from the shoreline and except near the Hudson Canyon, is a site where Taylor's analysis of dispersion in shear flow can be applied to any species resident in the water column for longer than 1 to 4 days, at least in winter. The application in summer is more difficult and will not be attempted here because of lack of adequate data; however it seems likely that during the period of the summer thermocline the shear flow analysis could be applied within layers.

Mixing caused by the tide—The tidal current on the shelf typically describes an ellipse with a major axis of about 15 $\mathrm{cm \cdot s^{-1}}$ maximum depth-averaged velocity parallel to the shore and 5 $\mathrm{cm \cdot s^{-1}}$ perpendicular to the shore (*see* Patchen et al. 1976). Velocity profiles have not been measured well enough to permit better than an approximate estimate of the mixing coefficient due to shear flow; I shall assume that the profiles are logarithmic as in Fig. 4C with $u' = 0.5\bar{u}$ and unidirectional at any instant, giving

$$D = (1/60)(\bar{u}^2 h^2/E) \qquad (6)$$

in the case of steady flow. I shall use Csanady's estimate of $E = 100\ \mathrm{cm^2 \cdot s^{-1}}$ for all subsequent calculations, with the understanding that all the computed mixing coefficients depend inversely on the value chosen for E and that Csanady's estimate is at best an approximation during unstratified conditions.

In the case of a very long period sinusoidal oscillation the shear flow mixing coefficient has a cycle-averaged value equal to half of what is obtained using the peak velocities (Fischer et al. 1979). On the shelf, values of TE/h^2 are not large; with $T = 12.5\ h$ we have values of 0.18 at 50-m depth and 0.045 at 100-m depth. Equation 6, with half of the peak values of $\bar{u} = 15\ \mathrm{cm \cdot s^{-1}}$ for long-shelf mixing and 5 $\mathrm{cm \cdot s^{-1}}$ for cross-shelf mixing, and with the result reduced by the factor taken from Fig. 5, predicts the values for a long-shelf mixing coefficient, D_l, and a cross-shelf mixing coefficient, D_c, given in Table 1. The table also gives standard deviations of a substance cloud, $\sigma = 2Dt$ where t is the time from release, for the case where the cloud has been dispersing for 5 days. σ_l is the long-shelf and σ_c the cross-shelf standard deviation. Both are less than the tidal excursion.

Table 1. Dispersion coefficients resulting from a 15–5 $\mathrm{cm \cdot s^{-1}}$ tidal ellipse and vertical mixing coefficient $E = 100\ \mathrm{cm^2 \cdot s^{-1}}$.

Depth, h(m)	50	100
$D_l(\mathrm{cm^2 \cdot s^{-1}})$	5×10^4	10^4
$D_c(\mathrm{cm^2 \cdot s^{-1}})$	5×10^3	10^3
σ_l(5 days) (km)	2	1
σ_c(5 days) (km)	0.7	0.3

Mixing caused by large storms—Large offshore storms often move up the Atlantic coast driving strong southwesterly currents. Boicourt and Hacker (1976) determined mean long-shelf velocity profiles at four stations off the mouth of Chesapeake Bay during a 59-h storm period in February 1974. Means for 59 h are given at three depths at the 35-m contour and also at the 60-m contour. At the 35-m contour, $\bar{u}$ (average of the three depths) = 33.6 $\mathrm{cm \cdot s^{-1}}$ and u' (maximum deviation from the average) = 4 $\mathrm{cm \cdot s^{-1}}$. At the 60-m contour, $\bar{u} = 40\ \mathrm{cm \cdot s^{-1}}$ and $u' = 4\ \mathrm{cm \cdot s^{-1}}$. The three profiles shown in Fig. 4 give $D = 1.3\times10^5$, 3.3×10^5, and 2.7×10^5 $\mathrm{cm^2 \cdot s^{-1}}$ at a depth of 50 m. During a storm the vertical mixing coefficient is likely to be higher than our assumed value of 100 $\mathrm{cm^2 \cdot s^{-1}}$, however, because of the higher wind stress. Not enough is known about the details of turbulence during a storm to be specific, but an estimate for the long-shelf mixing coefficient during the sort of large-scale persistent motion observed by Boicourt and Hacker would be of the order of 10^5 $cm^2 \cdot s^{-1}$. Hence the spread of an instantaneous source of substance, after a period of 2½ days of storm motion, would be of the order of 2 km, as compared to long-shelf advection by the mean velocity of the order of 75 km.

Boicourt and Hacker found that all the cross-shelf currents were landward during the storm period, except for one at the deepest station. For the cross-shelf component u' was of the order of 1 $cm \cdot s^{-1}$, suggesting that cross-shelf shear flow dispersion is negligibly small.

Mixing caused by the thermohaline circulation—Gradients of temperature and salinity drive a steady cross-shelf circulation which has been analyzed by Csanady (1976). His computed velocity profile is in reasonable agreement with observations summarized by Beardsley et al. (1976) and can be approximated by taking $u' = 5$ $cm \cdot s^{-1}$. The profile lies between the extremes of the linear gradient shown in Fig. 4A and the step of Fig. 4B, so the resulting mixing coefficient should also lie between the extremes computed from the two profiles, which are $D = 2.1 \times 10^5$ and 5.2×10^5 $cm^2 \cdot s^{-1}$ at the 50-m contour. A reasonable compromise prediction for the cross-shelf mixing coefficient due to the thermohaline circulation would be $D = 3 \times 10^5$ $cm^2 \cdot s^{-1}$.

Dispersion of substances

The distribution of short-lived substances is determined primarily by local currents rather than by turbulent diffusion and shear flow dispersion, because the spread of a cloud of particles within a few days is less than the tidal ellipse and much less than the distance the cloud may be carried by storm-driven currents. Observed distributions of short-lived tracers, such as radon-222, may be able to indicate rates of local mixing and short term advective transport, but not mixing on longer time scales such as the cross-shelf mixing caused by the thermohaline circulation. On the other hand, substances with lifetimes >100 days or so reside in the shelf flow long enough to average the periods of storm and calm weather, and to be affected by seasonal changes. To discuss the fate of these substances we need to consider the field of flow on the shelf as a whole, with and without a thermocline. Unfortunately the flow during the summer, which is characterized by a thermocline and strong stratification, reduced vertical mixing, and coastal jets and upwelling and downwelling, is extremely complex. Even if we confine our view to the midshelf region far enough from the coast to avoid coastal problems, the thermocline isolates the flow into at least two layers neither of whose motion is well enough understood to permit a quantitative analysis. Therefore in what follows we limit ourselves to wintertime, unstratified conditions.

Let us idealize the Middle Atlantic Bight as being a shelf 100 km wide (to the 100-m depth contour), of constant cross section and lying along a straight coast 1,000 km long (the approximate distance from Cape Cod to Cape Hatteras). The only significant inaccuracy in this picture is in the vicinity of the New York Bight and across the Hudson Canyon. In the apex of the New York Bight there appears to be a clockwise gyre of mean flow. The effect of the Hudson Canyon is not well documented, except that there appears to be a general bottom flow toward the mouth of the Hudson. Leaving these complications aside, the mean flow on the shelf appears to be in the range of 3–10 $cm \cdot s^{-1}$ to the southwest, the most frequently mentioned figure being 5 $cm \cdot s^{-1}$. This velocity corresponds to a mean travel time from Cape Cod to Cape Hatteras of 213 days, or about 7 months.

Suppose that a slug of long-lived material is deposited at the 50-m contour; what distribution can we predict 100 days later? At a velocity of 5 $cm \cdot s^{-1}$ the mean of the distribution will be advected along the shelf, approximately following the 50-m contour, a distance of about 430 km (unless, of course, it passes Cape Hatteras). If the long-shelf dispersion coefficient is 10^5 $cm^2 \cdot s^{-1}$ and the cross-shelf

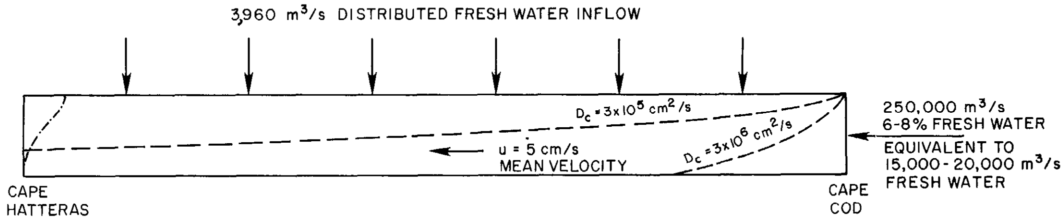

Fig. 6. Idealized view of yearly average flow in Middle Atlantic Bight. Dashed lines show edge (defined by 2 SD) of a plume of an inert tracer discharged at coast at northeast boundary, corresponding to cross-shelf dispersion coefficients of 3×10^5 and 3×10^6 $cm^2\cdot s^{-1}$. Dotted line at southwest boundary shows tracer distribution as flow approaches Cape Hatteras, corresponding to $D_c = 3\times10^5$ $cm^2\cdot s^{-1}$.

coefficient is 3×10^5 $cm^2\cdot s^{-1}$, the slug will disperse into an elliptical distribution having a cross-shelf standard deviation of about 23 km and a long-shelf standard deviation of 13 km. Since the distance from the 50-m contour to the shelf break is generally about 30 km, a small portion of the initial mass will reach the break and be exchanged across it.

Freshwater can be looked on as a long-lived tracer introduced into the saline water on the shelf, and the distribution of freshwater can be used as an indicator of mixing rates. Ketchum and Keen (1955) and Stommel and Leetmaa (1972) computed cross-shelf mixing coefficients by assuming that the flux of freshwater across the shelf is everywhere equal to the flux onto the shelf from tributary rivers. Ketchum and Keen computed values ranging from 0.6 to 5×10^6 $cm^2\cdot s^{-1}$ and Stommel and Leetmaa obtained a value of 3×10^6. The values given by both sets of investigators are substantially higher than that computed here. If my value is correct, however, the assumption used in the earlier computations is not valid, for the reason illustrated in Fig. 6. Yearly average freshwater inflows are shown on the figure: 3,920 $m^3\cdot s^{-1}$ comes from the rivers, primarily the Connecticut, Hudson, Delaware, and the Chesapeake Bay system (Bue 1970), and about 15,000–20,000 $m^3\cdot s^{-1}$ comes around Cape Cod. The latter estimate is based on the statement by Beardsley et al. (1976) that the flow around Cape Cod is about 250,000 $m^3\cdot s^{-1}$ and the finding of Brown and Beardsley (1978) that the salinity of this flow varies from 32.3 to 33‰, indicating that the flow is 6–8% freshwater. Figure 6 shows the nominal edge, defined as a distance of 2 SD, of a plume of tracer introduced at the Cape Cod end of the shelf diffusing outward into a flow with longshore velocity of 5 $cm\cdot s^{-1}$. If the cross-shelf mixing coefficient is 3×10^5 $cm^2\cdot s^{-1}$, the edge of the plume does not reach the shelf break before the flow reaches Cape Hatteras. Over most of the shelf, freshwater participating in cross-shelf exchange comes mostly from the flow around Cape Cod; therefore the cross-shelf flux bears hardly any relation to the inflow of freshwater from tributary rivers, and it is not possible to compute a cross-shelf mixing coefficient from the observed river discharge and cross-shelf salinity gradient.

We can hazard a very rough guess, based on the satellite and bathymetric data already reviewed, that at the shelf break there is an exchange flow of about 10^5 $m^3\cdot s^{-1}$ containing perhaps 2–4% freshwater. If so, this mechanism carries 2,000–5,000 $m^3\cdot s^{-1}$ of freshwater offshore; presumably most of it comes from the freshwater flow around Cape Cod. If this estimate is correct, the diffusion coefficients across the shelf break estimated by Ketchum and Keen and by Stommel and Leetmaa are of the right order of magnitude. The use of any diffusion coefficient at the shelf break is questionable, however; the offshore flux seems to take place mostly in large-scale gyres like the ones shown in Fig. 2, in which case the flux depends on the concentration in the water being carried offshore rather than on the local concentration gradient, and the process is more one of radiation than diffusion. Between the shelf break and the coast, on the other hand, the satellite pictures do not reveal significant large-scale motions; the Coast Guard maps occasionally suggest significant onshore or offshore displacements in local areas, but in winter the temperature contours are usually parallel to the coast. It seems likely that the magnitude of the cross-shelf mixing coefficient probably increases as one moves outward toward the shelf break, rising from the value I have estimated of 3×10^5 $cm^2 \cdot s^{-1}$ while at the same time the process gradually changes from one adequately described by a diffusion equation to one better described as radiation. Unfortunately the value I have estimated has not been confirmed by experiment; moreover, since all practical tracer experiments are likely to have the character of simulating the dispersion of short-lived tracers whose behavior we have seen to be quite different from long-lived ones, experimental confirmation seems difficult.

Toward a better understanding

The analysis sketched herein is preliminary and can be improved as more data become available. Improved knowledge of the rate of vertical mixing may be obtained through more comprehensive observations of radon distribution. More oceanographic studies can yield improved knowledge of currents, particularly a better definition of long term means and the shorter term variability during storms, and a better definition of the shear flow profiles. We can expect better estimates of the flux of freshwater around Cape Cod, past Cape Hatteras, and across the shelf break, and a better estimate of the frequency of impact of Gulf Stream rings. Besides more data, however, we need a better framework for analysis. Theoretical studies are unlikely to lead much further than the conclusions reported herein, but a simple numerical model based on the riverlike nature of the flow might be of considerable value. A coordinated program of field observations and numerical modeling ought to be able to answer the following questions, which I have answered only approximately or not at all.

What is the true rate of cross-shelf mixing, how does it vary across the shelf, and how is it affected by major storms?

What proportion of the freshwater exchange across the shelf break north of the mouth of Chesapeake Bay comes from rivers which discharge into the Middle Atlantic Bight, if any, and what proportion enters the bight in the flow around Cape Cod?

What is the distribution of freshwater entering the bight from individual rivers, such as the Connecticut and the Hudson, and how does it vary with the seasons?

Are the long-shelf and cross-shelf dispersion coefficients given here reasonable, or do larger scale motions induce much more mixing than we have estimated?

For that period of the year when the formation of the thermocline has an important effect on mixing, how can the effect be included in an analysis?

Is the rate of vertical mixing used herein of the right order of magnitude, and how is it affected by storms?

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