Simulating the Delaware Bay Buoyant Outflow: Comparison with Observations

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ABSTRACT

Coastal buoyant outflows from rivers and estuaries previously have been studied with field research, laboratory experiments, and numerical models. There is a dire need to evaluate model performance in light of coastal current observations. This research simulates the Delaware Bay outflow and compares results with observations of estuarine and shelf conditions. Observations include an estuarine salinity climatology, a record of freshwater delivery to the shelf, coastal current salinity mappings, and surface drifter data. Simulation efforts focus on spring 1993 and spring 1994, the primary field study period. The simulation is forced with river discharge, winds, and tides; only tidal-averaged results are discussed. Estuarine salinity results are consistent with the observed lateral salinity pattern, vertical structure, and response to river discharge. Salinities within the lower bay agree with observations, but the simulation overestimates the along-estuary salinity gradient. Observed and simulated freshwater delivery exhibit the same amplitude of response to river discharge and winds. The simulation produces a buoyant outflow that is generally consistent with the observed buoyancy signature, width, length, and vertical structure over a variety of river discharge and wind conditions. The simulated coastal current, however, tends to be somewhat shorter and fresher than observed. Simulated surface drifter paths exhibit the observed onshore advection during downwelling winds as well as offshore transport and current reversals during upwelling winds. A statistical evaluation based on shelf salinity mappings indicates that the model reproduces the observed variance and has only a small bias (less than 10% of plume buoyancy signature). The rms error of 1.2 psu is linked to the shorter and fresher nature of the simulated coastal current. Observational comparisons discussed in this paper indicate that the model can simulate many coastal current features and its response to river discharge and wind forcing.

1. Introduction

A top priority in coastal oceanography is to understand the flow and mixing of buoyant waters and terrestrial materials into the coastal ocean (Brink et al. 1992). Freshwater inputs from rivers and estuaries fuel coastal buoyant outflows on continental shelves. These coastal currents have been studied with field observations, laboratory experiments, and numerical simulations. Despite the frequent use of hydrodynamic models, there is a dearth of published comparisons between simulation results and coastal current observations. The call made by the National Research Council (1993) for systematic evaluation of simulation results in order to “insure the scientific integrity of modeling products” largely has gone unanswered.

Instead of conducting quantitative observational comparisons, most coastal current modeling studies use observations only to motivate the research or identify some general features reproduced. In some cases, the availability of observations is limiting. For example, Kourafalou et al. (1996b) simulated the buoyant outflow onto the South Atlantic Bight, but conclude “survey data do not offer a conclusive validation of model results.” In other cases, few comparisons are made despite abundant observations. For instance, García Berdeal et al. (2002) model situations representative of the Columbia River plume but miss an opportunity to quantitatively compare results with copious data collected by Hickey et al. (1998).

The objective of this paper is to evaluate model performance in light of field observations; this represents the first such effort for coastal buoyant outflows. The case study discussed simulates the freshwater outflow from the Delaware Bay into the Middle Atlantic Bight. A wealth of data from the Delaware Bay and Delaware coastal current allows a comprehensive assessment of...
model performance. The simulation is run without data assimilation; observations are used only for comparison purposes.

The following section describes the study area and physical characteristics of the Delaware Bay buoyant outflow. Sections 3 and 4 detail methods used to configure the simulation. Section 5 describes the observational data used for model evaluation. Section 6 compares results with estuarine salinity observations. Section 7 discusses simulated and observed freshwater delivery to the shelf. Section 8 compares results with shelf salinity data taken during shipboard mappings. Section 9 adopts a Lagrangian viewpoint to assess how well simulated drifters match actual drifter data. The final section summarizes findings and discusses implications for other numerical studies of coastal buoyant outflows.

2. Study area

Tidal portions of the Delaware River (beginning at Trenton, New Jersey) and the Delaware Bay constitute the Delaware estuary, a major estuary on the U.S. East Coast. This coastal plain estuary stretches 210 km to the mouth, draining a land area of $4.2 \times 10^5$ km$^2$ (Sharp 1984). The bay widens down-estuary from the river; maximum bay width is 45 km. The bay narrows to 18 km at the mouth between Cape Henlopen and Cape May. Mean estuary depth is 7 m (Harleman 1966). The deepest waters (exceeding 30 m) are associated with the ancestral channel that cuts a path through the mouth and onto the shelf near the Delaware coast (Garvine 1991).

The Delaware River above the estuary head provides 58% of the freshwater inflow. The confluence of the Schuylkill River below Philadelphia adds another 14%. Other sources collectively account for the remaining 28%; each contributes no more than 1% individually (Sharp 1984). Average freshwater inflow is 570 m$^3$ s$^{-1}$. High river discharge conditions occur during the spring. The typical river discharge rate during April is 1100 m$^3$ s$^{-1}$. Even under peak freshwater inflow, the estuary is vertically well mixed by the tides (Wong 1995). Brackish waters can intrude as far as 125 km from the mouth up to Chester, Pennsylvania (Sharp 1988).

Spring river discharge conditions deliver much freshwater to the shelf. Upon exiting the mouth, buoyant waters turn anticyclonically under the influence of the earth’s rotation. During light winds, the buoyant outflow propagates downshelf (defined as the direction of Kelvin wave propagation) along the Delmarva Peninsula coast in a slender coastal current. On the inner shelf, winds drive across-shelf flow through Ekman transport and geostrophic alongshelf flow (Clarke and Brink 1985). The wind-driven alongshelf flow is barotropic along the Middle Atlantic Bight (e.g., Chuang et al. 1979; Yankovsky and Garvine 1998; Münchow and Chant 2000). Downwelling-favorable winds tend to accelerate downshelf flow in the coastal current and compress its waters against the coast. Upwelling-favorable winds counter buoyancy-driven downshelf flow and tend to spread buoyant waters offshore (Whitney 2003).

The Delaware coastal current belongs to the class of large-scale buoyant outflows since its internal Kelvin number (ratio of current width to internal Rossby radius, $L_c/R$) exceeds 1 (Garvine 1995). The lowest-order across-shelf momentum balance is geostrophic; an across-shelf pressure gradient supports the buoyancy-driven alongshelf current. Scaling of the alongshelf momentum equation reveals that the Coriolis term involving the across-shelf flow can be balanced by the alongshelf pressure gradient, wind stress, and bottom stress (Garvine 1995).

The shallow bathymetry of the gently sloped inner shelf causes the Delaware coastal current to interact extensively with the bottom. This contact influences coastal current dynamics and leads to bottom trapping—a process that advects the foot of the plume out to the scale depth imposed by thermal wind dynamics (Chapman and Lentz 1994). The location of this trapping depth causes the core of the coastal jet to be several kilometers offshore (Yankovsky and Chapman 1997).

3. Model configuration

The simulation is conducted with the three-dimensional Estuarine, Coastal, and Ocean Model (ECOM3d). It is closely related to the Princeton Ocean Model (Blumberg and Mellor 1983, 1987), a standard in the field. Governing equations and their finite-difference form are given in the ECOM3d manual (Blumberg and Mellor 1995) and in Kourafalou et al. (1996a). Vertical resolution is provided by bathymetry-following sigma coordinates.

a. Model domain

The simulation domain (Fig. 1) includes the entire Delaware estuary and 340 km of the adjacent continental shelf (extending out to the 100-m isobath). The river course has been artificially bent in the model (apparent in Fig. 1) to keep the tidal reaches of the Delaware River within the smallest possible grid. Testing shows that these river course changes only have local effects. The included section of continental shelf stretches from
the mouth 110 km upshelf (opposite the direction of Kelvin wave propagation) past Atlantic City, New Jersey, and 230 km downshelf (to the south) along the Delmarva Peninsula to the Chesapeake Bay mouth. The domain is designed to focus on the Delaware buoyant outflow and to keep it within the grid.

The model grid contains 300 × 150 cells covering the 340 km × 240 km domain. Highest horizontal resolution (0.75–1 km) is required near the bay mouth and within the inner shelf, the primary coastal current region. The internal Rossby radius (approximately 6 km) is resolved wherever the buoyant plume reaches. Larger grid cells (up to 8 km wide across-shelf) can be tolerated far offshore. Vertical resolution is provided by 15 sigma levels; spacing is 5% of the water column near the surface and bottom (upper and lower 5 levels) and 10% in the interior. This vertical spacing allows for computational economy while still resolving features of interest.

Figure 1 includes contours of model bathymetry. Coastline and bathymetry (on the shelf and within the bay) are based on data provided by the NOAA National Geophysical Data Center. In the Delaware River, the model relies on river width and depth information published by Thatcher and Harleman (1981). Minimum water depth within the model grid is 3 m. The Chesapeake Bay and many tidal inlets lie within the model domain but are considered to be land in the simulation in order to focus on the Delaware Bay outflow.

b. Standard settings

ECOM3d uses a split time step to solve the depth-averaged and vertically varying fields. Temporal differencing for the simulation is set at 9.2 s for the depth-averaged part and 92 s for the depth-resolving part to avoid numerical instabilities. ECOM3d has been implemented with a nonlinear set of governing equations including momentum advection terms. The model is run with a recursive salt advection scheme (Smolarkiewicz 1983). Bottom friction follows the quadratic drag law. The bottom friction coefficient $C_D$ has been set to $5 \times$
$10^{-3}$ (2 times the standard value) to best match model results with observed tidal characteristics within the estuary. Bottom roughness is parameterized with a standard height of 0.3 cm.

Horizontal turbulent closure is accomplished with the Smagorinsky diffusion formula (Smagorinsky 1963). The empirical coefficient is set to 0.1, the same value used by Kourafalou et al. (1996a,b) and Garvine (1999). The horizontal turbulent Prandtl number is set at 1. The model combines the Mellor–Yamada (level 2.5) vertical turbulent closure scheme (Mellor and Yamada 1982) with a background viscosity level. The background vertical viscosity and diffusivity both are set at $2 \times 10^{-5}$ m$^2$ s$^{-1}$. This value is in the middle of the range tested by Garvine (1999). The model is initialized with homogeneous water (10°C, 32.0 psu); there is no ambient shelf stratification. Density depends solely on salinity since the simulation has no temperature variations.

c. Boundary conditions

Momentum flux at the free surface is imparted by wind stress. Momentum flux at the bottom boundary is balanced by bottom stress. A quadratic drag law is followed at both boundaries. Salt and heat flux are set to zero at the surface, bottom, and land boundaries. A partial-slip condition (standard to the model) is applied along land boundaries, setting tangential velocity at the coast to one-half of the value at a distance of one-half of a grid cell offshore. Varying this boundary condition from no slip to free slip has little effect on coastal current dynamics because the buoyancy-driven current is displaced offshore by bottom trapping.

The free surface at the offshelf open boundary is fixed, or “clamped,” to the tides by specifying tidal amplitude and phase. A “smooth” condition is applied for the tangential velocity (alongshelf flow): its boundary-normal gradient is set to zero. Salinity also has a smooth condition imposed along the offshelf side. This salinity boundary condition is unimportant since this boundary is intentionally far removed from coastal buoyant waters.

The free surface at the across-shelf boundaries must be able to oscillate with tides, set up with winds, and fluctuate with passing waves. Neither purely clamped nor purely passive conditions work well for the combination of forcings and model bathymetry. The remedy is a composite scheme that clamps only part of the surface elevation to known tidal information ($\eta_T$).

It is assumed that the tidal solution has no alongshelf (boundary normal) variation at the upshelf and downshelf boundaries. The residual elevation ($\eta - \eta_T$) has a nonzero alongshelf gradient and is handled with a passive wave radiation condition. Surface elevation at the boundary is found by finite-differencing the following equations ($c_w$ is the phase speed of an outgoing wave):

\[ \eta_T = \sum_i \eta \cos(\omega t - \varphi_i) \quad \text{for } i \text{ constituents} \]

\[ \frac{\partial \eta}{\partial t} = \frac{\partial \eta_T}{\partial t} + c_w \frac{\partial \eta}{\partial x}. \]

The composite scheme applied combines the clamped and implicit gravity wave radiation ($c_w = \sqrt{gh}$) conditions tested by Chapman (1985).

River discharge is imposed as a vertically uniform inflow at the head of the estuary. Without this river input, there would be no volume flux at this location. No special condition is required for the free surface at this location.

4. Model forcing

Three primary forcing agents are included: Delaware River discharge, winds, and tides. The simulation is forced as simply as possible while still capturing the essence of the Delaware coastal current system. In the simulation, sea level variations and currents are driven exclusively by the tides, local winds, and the buoyant outflow. Consequently, the model does not simulate the mean flow that slowly transports shelf waters downshelf from Cape Cod to Cape Hatteras (Beardsley and Winant 1979). Remote wind events can generate barotropic shelf waves that propagate downshelf through the study area (at a rate of 930 km day$^{-1}$) and modulate shelf currents as they pass (Yankovsky and Garvine 1998). The slowly varying large-scale shelf circulation and the transient currents associated with shelf waves are excluded from the simulation. The focus is on the response to local winds that dominates the variability in the weather-band frequencies. There is no ambient shelf stratification in the standard runs. During the spring study period, seasonal stratification is weak; waters along the inner shelf remain thermally mixed through May (Kohut et al. 2004). Furthermore, the typical stratification associated with the plume (Münchow and Garvine 1993a) is an order of magnitude larger than ambient shelf stratification along the Middle Atlantic Bight (Clarke and Brink 1985) and the wind-driven alongshelf currents are barotropic throughout the year (Münchow and Chant 2000). The Hudson and Chesapeake plumes enter the shelf to the north and south of Delaware Bay but are not included in the simulation. This exclusion is justified since the Hudson, Delaware, and Chesapeake plumes remain separated (Münchow and Garvine 1993b) because of the distance between their inflow locations (greater
than 200 km). Consequently, it is appropriate to include only the Delaware buoyant outflow in the simulation.

The simulation begins with a period of tidal forcing. Following this tidal spinup period, river discharge and winds are imposed. Efforts focus on the spring discharge period of 1993. The 1993 run starts discharge and winds on 26 December 1992 and ends on 1 July 1993. A run starting one year later was completed for comparison with spring 1994 observations.

a. Tides

Tidal mixing plays a crucial role within the estuary and immediately outside the bay mouth. Consequently, tides must be included. Simulations are forced with the dominant tidal constituent: the semidiurnal lunar tide ($M_2$). Tidal heights are imposed along open boundaries by specifying amplitude and phase. Values specified along the offshore boundary (100-m isobath) come from results of an inverse tidal model incorporating Ocean Topography Experiment (TOPEX)/Poseidon altimetry data (Egbert et al. 1994). Within the study area, tidal amplitude along the shelfbreak is a uniform 44 cm and phase increases slightly upshelf by less than $2^\circ$.

Boundary conditions demand tidal information along all open boundaries, as in the shelf tidal modeling efforts of Han (2000). The inverse tidal model does not provide reliable results along the across-shelf boundaries because TOPEX/Poseidon altimetry does not perform well over shallow waters. To create tidal information for the across-shelf open boundary conditions, a tidal simulation over a much longer domain (with lower resolution and no estuary) was executed prior to the tidal simulation for the across-shelf open boundary conditions, a tidal simulation. Consequently, it is appropriate to include only the Delaware buoyant outflow in the simulation. Efforts focus on the spring discharge period of 1993. The 1993 run starts discharge and winds on 26 December 1992 and ends on 1 July 1993. A run starting one year later was completed for comparison with spring 1994 observations.

b. River discharge

River discharge constitutes the most important forcing since there would be no estuary or buoyant outflow without freshwater input. The Delaware River above Trenton, New Jersey, provides 58% of the freshwater flowing into the Delaware Bay (Sharp 1984). The U. S. Geological Service daily river discharge record at Trenton is used to force the simulation. These values are stepped up accordingly to represent the entire freshwater inflow and are applied at the estuary head.

River discharge is introduced into the model after 62.1 days (120 tidal cycles). At this time, an estuary salinity field also is imposed. This allows the simulated estuary salinity field to quickly converge upon a quasi-steady state. This imposed field is based on the alongestuary salinity trend empirically derived by Garvine et al. (1992). It has a linear axial salinity distribution, no across-estuary gradient, and is vertically homogeneous upon initialization. This salinity field is subsequently mixed by the tides. The estuary reaches quasi-steady state within several weeks. In section 6, comparisons are made between simulation results and the same observed salinity trend used to initialize the estuarine salinity field. These comparisons are valid because enough time has passed to allow the simulated salinity field to evolve to a state that is different from the initialization field. By the comparison period, estuarine salinities are the same as those that eventually develop in longer runs without an imposed salinity field.

River discharge is initiated three months before the onset of the spring discharge period (on 26 December 1992 for the 1993 run). This standard run produces the same results as a test run initiated five months prior to spring discharge; whereas a test run begun only a month prior exhibited different results during the early spring. Thus, it takes over a month for the buoyant outflow to grow independent of initialization conditions.

Spring 1993 is wetter than usual; April 1993 monthly averaged river discharge (2400 m$^3$ s$^{-1}$) is 2 times the long-term average value for April. Consequently, the standard simulation period exhibits a strong outflow and is well suited for studying the Delaware coastal current. Figure 2a shows river discharge forcing for 1993. An intense pulse of river discharge (over 5000 m$^3$ s$^{-1}$) occurs at the beginning of April 1993. Spring 1994 has similar monthly-averaged discharge (not shown) but is characterized by more evenly distributed freshwater inflow.

c. Winds

Time-variable winds are imposed as a spatially uniform forcing representative of inner shelf conditions. Excluding spatial variability in the wind field is acceptable since midlatitude weather systems typically are larger than the model domain. Winds from all directions are included in the simulation even though shelf wind response is dominated by alongshelf winds (Mitchum and Clarke 1986). Hourly surface meteorological observations for the Atlantic City International Airport (in Atlantic City, New Jersey) were obtained from the Northeast Regional Climate Center and 40-h low-pass filtered. Wind data from the weather buoys anchored offshore of Delaware Bay would have been preferable, but long data gaps occurred during spring
estuary and on the shelf. An estuarine salinity climatology has been assembled from 822 conductivity temperature and depth (CTD) profiles stored in a National Oceanographic Data Center (NODC) database. These data (collected during all seasons and spanning from 1950 to 2000) have been averaged together to produce a salinity field representative of mean conditions. An along-estuary salinity trend has been extracted from this climatology. Another source for the axial salinity trend comes from several long-term instrument records analyzed by Garvine et al. (1992).

Simulation results on the shelf are compared with observations from the Delaware Coastal Current Experiment field study, conducted during the 1993 and 1994 high-discharge seasons (Sanders and Garvine 2001). Moorings were deployed along an arc outside the bay mouth (Fig. 1 inset) to capture buoyant outflow source conditions. The moored instruments recorded salinity and current data throughout spring 1993. Shipboard surveys mapped the distribution of buoyant waters on the shelf. Drifters were deployed to provide Lagrangian trajectories and additional salinity information. Four surveys and drifter deployments completed in 1993 complement observations provided by the moored instruments. The second year of the study consisted of three shipboard surveys and drifter mappings. Observations have been analyzed in Sanders (1999) and Sanders and Garvine (2001).

A detailed comparison between simulated and observed $M_2$ tides is included in Whitney (2003); here it is summarized only briefly to focus on subtidal results. The simulation reproduces the observed across-shelf amplification of tidal heights. Tidal elevation amplitudes agree within 2% along the shelf and within 5% in the bay. Simulated and observed shelf tidal currents have similar amplitudes (around 10 cm s$^{-1}$) and exhibit the same rotary nature, with major axes oriented across-shelf and an ellipticity of $-0.7$. Simulated results are consistent with the observed up-estuary increase in tidal phase. Tidal current amplitudes within the lower bay (around 60 cm s$^{-1}$) are overestimated by 10 cm s$^{-1}$. Consequently, the tidal volume flux through the mouth ($1.7 \times 10^7$ m$^3$ s$^{-1}$) is 16% larger than observed ($1.5 \times 10^7$ m$^3$ s$^{-1}$).

Linear regressions are used to compare simulation results with observations. Every reported correlation coefficient $r$ is significantly different from zero at the 95% confidence level. Each correlation coefficient and regression slope $b$ is shown with its 95% confidence interval. These confidence intervals are computed following the methods described in Emery and Thomson (2001) with the effective degrees of freedom ($N_{\text{eff}}$) replacing the number of data points ($N$). Here $N_{\text{eff}}$ is
calculated with the formula in Chelton (1983); it provides a conservative estimate that accounts for the auto- and cross-correlations of the data series. As in Pineda and Lopez (2002), the maximum lag used for the $N_{eff}$ calculations is 20% of the series length.

6. Estuary salinity

This section compares results within the estuary with long-term salinity observations. Reproducing the horizontal salinity distribution and vertical stratification is important within the lower bay, as this provides source conditions for the Delaware coastal current. Figure 3 compares the 1993 average salinity field based on simulation results with the climatology constructed from the NODC database. NODC depth-averaged salinities are indicated by color-coded numbers and simulation results are color contoured. Simulated and observed salinity fields exhibit the same pronounced across-estuary structure. Low salinities extend farther down-estuary in the shallow regions along each coast than in the bay center. The shallow area (less than 3 m deep) behind Cape May contains salinities below 20 psu. This pattern of freshwater on the shallow flanks is a persistent feature throughout the simulation run. Wong (1994) attributes this pattern to the nature of gravitational circulation over laterally variable bathymetry. Applying a momentum balance between the axial pressure gradient and vertical shear stress over a triangular cross section, Wong (1994) shows dense water inflow is centered over the deepest section and freshwater outflow is situated over each flank. Kasai et al. (2000) show that this structure develops because high vertical eddy viscosity and shallow depths lead to a large Ekman number, eliminating Coriolis acceleration from the lowest-order momentum balance.

Along-estuary salinity trends have been extracted from the NODC climatology and the 1993 average simulation results; these are shown in Fig. 4a. This figure also includes the trend line and salinity observations published in Garvine et al. (1992). The climatology follows the trend line slope of $1/3$ psu km$^{-1}$ along most of the bay. Near the mouth and in the river, the climatology reveals smaller salinity gradients (less than $1/8$ psu km$^{-1}$) that the Garvine et al. (1992) data cannot resolve. The simulated estuary is characterized by a shorter transition from ocean water to freshwater. Some of this difference arises from comparing simulation results representative of 1993 (a wet year) with long-term average conditions. The 1993 record has an average river discharge of 640 m$^3$ s$^{-1}$ and a maximum monthly averaged discharge of 2400 m$^3$ s$^{-1}$ (during April). In comparison, the 1950–2000 NODC period has a lower average discharge of 570 m$^3$ s$^{-1}$ and a maximum monthly-averaged discharge of 1100 m$^3$ s$^{-1}$ (during April).

Salinity at every location is below the observed trend; this difference intensifies up estuary. Beyond 40 km from the mouth, the axial salinity curve of the simulation is more than one standard deviation away from observations. The maximum along-estuary salinity gradient is double the observed value. The higher salinity gradient and subsequent shorter distance between fresh and ocean water points to less mixing in the model. Lower mixing levels may be due to the across-estuary coarseness of the model grid. The lower bay and lower

Fig. 3. Delaware Bay average salinity field. (a) NODC climatology; colored numbers indicate depth-averaged salinity values. Data have been binned in 5-km square cells. (b) Depth-averaged simulation results averaged over 1993.

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river are characterized by low salinity gradients similar to the climatology. The average axial salinity trend in the simulation is similar to that of an earlier modeling effort by Galperin and Mellor (1990c). The earlier simulation, however, exhibits strong response to river discharge not seen in observations or the present simulation.

The salt balance in its cross-sectionally and tidally averaged form (Monismith et al. 2002) is

$$A \frac{dS}{dt} + \frac{dQS}{dy} = \frac{d}{dy} \left( K_{sy} A \frac{dS}{dy} \right).$$  \hspace{1cm} (2)

For the Delaware, the cross-sectional area \(A\) increases with along-estuary distance \((y)\). Salinity \(S\) does not change with distance in the lower river and bay. Considering a steady-state balance with spatially constant eddy diffusivity \(K_{sy}\), for simplicity yields

$$Q \frac{dS}{dy} = K_{sy} \frac{dA}{dy} \frac{dS}{dy} + K_{sy} A \frac{d^2S}{dy^2}. \hspace{1cm} (3)$$

The first term, representing down-estuary salt advection by freshwater inflow, is balanced by the dispersive salt transport up estuary (Monismith et al. 2002). NODC observations and simulation results both indicate that the second term in Eq. (3) is larger than the final term over most of Delaware Bay (farther than 30 km from the mouth). Neglecting the final term leads to a two-term steady state balance:

$$Q = K_{sy} \frac{dA}{dy}. \hspace{1cm} (4)$$

Over most of Delaware Bay, geometrical dispersion (lateral mixing) generated by the down-estuary widening of the bay balances discharge-driven salt advection. Equation (4) implies that increased discharge leads to a higher eddy diffusivity. This buffering leads to a weak dependence between estuarine salinity and river discharge. Garvine et al. (1992) investigated this dependence using long-term salinity records and discharge data. At each station, salinity is negatively correlated with \(Q\) (significant at the 95% confidence level), indicating decreased salinity with increased discharge. The regression slope \(dS/dQ\) for each station is included in Table 1. Change in isohaline location \((y_s)\) with discharge was estimated by dividing \(dS/dQ\) by the mean axial salinity gradient of 1/3 psu km\(^{-1}\); \(dy_s/dQ\) is included in Table 1. This analysis reveals the weak dependence between salinity and river discharge anticipated by Eq. (4). A one-standard-deviation change in discharge (approximately 330 m\(^3\) s\(^{-1}\)) shifts salinities between 0.5 and 2.6 psu. This displaces isohalines by at most 8 km, which is less than the 9-km excursion produced by the tides (Garvine et al. 1992).

Simulation results exhibit the expected negative correlation (significant at the 95% confidence level) between salinity and discharge: regression slopes at each station are included in Table 1. These regression slopes have been translated into changes in isohaline location \((dy_s/dQ)\) using the axial salinity trend of the 1993 average simulation results (Fig. 4a). Observed and simulated regression slopes are of the same order of magnitude. In the lower bay, \(dS/dQ\) values indicate one-half
of the observed dependence. The isohaline shift (5–7 km per $10^3$ m$^3$ s$^{-1}$), however, matches observations because of the gentler lower-bay salinity gradient in the model (and in the NODC climatology). Farther up-estuary the isohaline shift is smaller than observed, 0–5 km per $10^3$ m$^3$ s$^{-1}$ as compared with 12–23 km per $10^3$ m$^3$ s$^{-1}$. The simulation results agree with the observed response to river discharge much better than the earlier modeling efforts of Galperin and Mellor (1990c). Their results (with a $dy/dQ$ of 60 km per $10^3$ m$^3$ s$^{-1}$) are far too sensitive to river discharge.

Figure 4b graphs the along-estuary trend in vertical salinity range for the NODC climatology and the 1993 average simulation results (along the estuary thalweg). The curves have similar shapes; however, the larger axial salinity gradient in the simulation causes entirely freshwater (with no vertical salinity range) to be located closer to the mouth than observed. Each curve has a maximum near the same position (40 km from the mouth). The observed maximum vertical salinity range is 3.8 psu. The maximum from simulation results (2.9 psu) is less than the observed value but is well within the standard deviation envelope bounding the NODC curve. This comparison indicates that simulation results are consistent with the tidally well-mixed nature of the Delaware estuary.

7. Freshwater delivery

During the Delaware Coastal Current Experiment, instruments were deployed along an arc (Fig. 1 inset) located 15 km downshelf of the mouth. This mooring arc captured freshwater delivery to the shelf. Salinity and current information were combined to calculate freshwater flux ($F$) through the vertical cross-sectional area of the mooring arc:

$$F = \int \left( S_u - \frac{S}{S_u} u \right) da.$$  (5)

In Eq. (5), $u$ represents velocity normal to the cross-sectional area ($da$ is the area element) and $S_u$ is the salinity of ambient shelf water. Sanders and Garvine (2001) calculated the freshwater flux time series shown in Fig. 5 from low-pass filtered (using a 40-h window) salinity and current records and an ambient salinity of 31.9 psu. Lagged river discharge also is plotted for reference; it has been lagged by five days to adjust for signal travel time through the estuary (following Sanders and Garvine 2001).

Observed freshwater flux ranges between 0 and 2700 m$^3$ s$^{-1}$. During the same period, river discharge varies between 200 and 5200 m$^3$ s$^{-1}$. Freshwater flux through the mooring arc responds to river discharge at low frequencies; the correlation coefficient is 0.62 (0.18 < $r$ < 0.85; $N_{eff} = 16$) (Sanders and Garvine 2001). The moored array captures most of the freshwater delivered to the shelf. Over the study period, the net freshwater volume passing through the arc accounts for 90% of the river inflow. The freshwater flux record does not follow the extreme peak in the $Q$ time series. Response to this intense discharge event is spread over a longer period; changes within the estuary buffer conditions on the shelf. Since freshwater can be stored within the estuary, freshwater flux through the arc is lower than river discharge in early spring and greater than river discharge in late spring.

The freshwater flux time series also exhibits shorter-period oscillations that are linked to wind influence. Pronounced variations between 15 and 27 April are linked to three strong wind events (Fig. 2b). Sanders and Garvine (2001) found a correlation of $r = 0.57$ (0.32 < $r$ < 0.74; $N_{eff} = 42$) for winds aligned 140°/320° and lagged by 12 h. This wind alignment produced the maximum response; it is rotated 70° counterclockwise from the large-scale shelf orientation and 20° clockwise from the mouth. The wind response reflects competition between shelf influence and estuarine effects.

Freshwater flux time series can be calculated from simulation results. Freshwater flux through grid cell sides is calculated at each time step using instantaneous salinity and velocity values [following Eq. (5)]. Ambient salinity is set at 32.0 psu. The time series included in Fig. 5 represents tidal-averaged freshwater flux through the arc shown in the Fig. 1 inset. The simulated and observed time series are similar except during the period between 27 April and 14 May when simulation results are consistently higher. Simulation results, like
the observed freshwater flux, show a response to river discharge and wind forcing. The freshwater flux time series correlates well to the observed record; the maximum correlation of \( r = 0.73 \) (\( 0.36 < r < 0.90 \); \( N_{\text{eff}} = 16 \)) occurs at zero lag. The linear regression between observations and model results reveals a regression slope of \( 0.90 \pm 0.47 \). The wind response of freshwater delivery agrees well with observed behavior: the wind alignment producing maximum response and the amplitude of wind-induced oscillations are the same as observed. Simulation results are consistent with the observed freshwater flux passing through the source region of the Delaware coastal current.

8. Coastal current surface salinity and vertical structure

Much insight concerning simulation performance and the nature of the buoyant outflow can be gained by looking at shelf salinity fields. Seven shipboard surveys mapped the Delaware coastal current during spring 1993 and spring 1994. Surface salinity and temperature data were collected en route by continuously pumping water from 0.5-m depth into a thermosalinograph system. The 1993 observed surface salinity fields are contoured in the upper panels of Fig. 6 (data transects also are shown). Lower panels contour surface salinity fields based on tidal-averaged simulation results corresponding to each mapping period. Comparing these fields indicates how well plume buoyancy signature, width, and length are simulated.

CTD profiles collected along ship transects allow for a comparison between observed and simulated vertical salinity structure. Of particular interest is how well the simulation reproduces the slopes of isohalines (coincident with isopycnals) and bottom contact of the buoyant outflow for different river discharge and wind conditions. A cross section 30 km downshelf is selected for comparison purposes because it closely follows a CTD line for each survey. Figure 7 presents the salinity cross sections. Observations are contoured in the upper panels and corresponding simulation results are contoured in the lower panels. The contour range (from 27 to 32 psu) is one-half of the range used in Fig. 6 in order to highlight the salinity structure at this coastal current section.

Tables 2 and 3 list plume characteristics derived from the data presented in Figs. 6 and 7. Table 2 includes the maximum salinity anomaly (using \( S_a = 32.0 \) psu), width...
(following the 31-psu contour), and Kelvin number range (using $R = 6$ km) for each mapping. Table 3 lists the downshelf extent of the 31- and 29-psu isohalines. This table reports the degree of bottom contact at the Fig. 7 cross section. Bottom contact is quantified by dividing the offshore distance of the 31-psu isohalines at the bottom ($Y_b$) by its offshore distance at the surface ($Y_s$). This ratio ($Y_b/Y_s$) varies between 0 for no bottom contact and 1 for full contact.

a. **DCC1 mapping: Onset of high river discharge, after downwelling winds**

The DCC1 mapping (3–5 April 1993) presents a snapshot of a growing buoyancy-driven outflow developing in response to the onset of high river discharge rates. Lagged river discharge (Fig. 5) grew from 2300 to 3700 m$^3$/s during the survey. Prior to the mapping, there was a two-day period of downwelling-favorable winds: peak winds 7 m s$^{-1}$ downshelf (Fig. 2b). During the survey, winds were light (below 2 m s$^{-1}$).

Observed surface salinities (Fig. 6a) range between 25.4 and 32.0 psu. Minimum salinities occur closest to the mouth; here the salinity anomaly is 6.6 psu. Plume width varies between 14 and 43 km. The coastal current Kelvin number ranges from 2.3 to 7.2; it is near 3 over most of the plume. Buoyant waters stretch far downshelf: the 31- and 29-psu isohalines penetrate 190 km and 130 km downshelf, respectively. Corresponding simulation results are shown in Fig. 6e. The simulation produces a plume with a stronger buoyancy signature than observed; the maximum salinity anomaly is 3.0 psu larger. Plume width varies between 10 and 20 km ($K$ is between 1.6 and 3.3). Typical plume width ($K = 3$) is similar to observations, but plume length is shorter. The 31- and 29-psu contours penetrate about half the observed distance. The 28-psu contour, however, penetrates 20% farther than observed to 50 km downshelf. At this time, the simulation produces a fresher outflow extending 50 km downshelf. Meanwhile, buoyant waters farther downshelf (delivered to the shelf during

| Table 2. Maximum salinity anomaly and range in plume width. Observed and simulated values are derived from surface salinity fields contoured in Fig. 6. Plume width is based on the offshore distance of the 31-psu isohaline, and Kelvin number values are based on an internal Rossby radius of 6 km. |
|-----------------|-----------|-----------|-----------|
| DCC1            | 6.6       | 14–45     | 2.3–7.2   |
| DCC2a           | 5.2       | 7–25      | 1.2–4.2   |
| DCC3            | 8.2       | 10–34     | 1.7–5.7   |
| DCC4            | 3.5       | >60       | >10       |
lower discharge and advected downshelf by downwelling winds) are much saltier than observed.

Observed and simulated vertical structure of the buoyant outflow are compared in Fig. 7. The observed plume (Fig. 7a) possesses steep isohalines with a characteristic slope of $7 \times 10^{-3}$. Downwelling winds prior to the survey excited Ekman transport (onshore at the surface and offshore near the bottom) that steepened the isohalines. At this time, there is extensive bottom contact ($Y_s/Y_r$ is 0.7). The simulated plume (Fig. 7e) has the same degree of bottom contact and isohalines have been steepened by downwelling winds. However, isohaline slopes ($2 \times 10^{-3}$ to $3 \times 10^{-3}$) are less than one-half of the measured slope. The gentler slopes and larger salinity anomaly indicate higher stratification than observed.

There are conspicuous differences between the simulated and observed plumes. During the DCC1 survey, the coastal current is responding to an intense river discharge pulse that switches the buoyant outflow from a low discharge to a high discharge regime. The simulation does not reproduce the exact timing or observed character of this transition to the high discharge state. Nevertheless, many of the general features are consistent with the observed field: the model produces a slender buoyancy-driven coastal current approximately three Rossby radii wide that propagates downshelf in contact with the bottom. Unlike in many model studies with only buoyancy forcing (e.g., Oey and Mellor 1993; Chapman and Lentz 1994; Yankovsky et al. 2001), artifacts such as upstream propagation and a large recirculation region outside the mouth are not present.

### b. DCC2a mapping: High river discharge, downwelling winds

The DCC2a mapping (29–30 April 1993) presents a snapshot of the coastal current under high river discharge conditions ($1700 \text{ m}^3 \text{ s}^{-1}$). Three days prior to the mapping, strong upwelling winds reversed the coastal current and mixed away the buoyant waters. The mapping coincides with the end of a downwelling-favorable wind event. Winds blew downshelf for four days with peak speeds of $8 \text{ m s}^{-1}$ occurring on 27 April (Fig. 2b). This wind event was strong enough to influence the coastal current by increasing downshelf transport, compressing buoyant waters against the coast, and decreasing stratification.

Figure 6b contours the observed surface salinity field. The maximum salinity anomaly of 5.2 psu is not as strong as in the DCC1 mapping. Plume width (7–25 km) is narrower because of onshore surface Ekman transport excited by the downwelling-favorable winds. Plume length is shorter than observed during the DCC1 mapping because the buoyant outflow was mixed away by upwelling winds a few days prior. Corresponding simulation results are shown in Fig. 6f. The maximum salinity anomaly (5.8 psu) is only 12% higher than observed. Like the observations, the simulated plume is narrower (8–18 km) than it was during the DCC1 time period. The simulated plume remains shorter than observed: the 31- and 29-psu isohalines penetrate to 60% and 73% of their observed extent, respectively.

Like the DCC1 mapping, the observed coastal current (Fig. 7b) has steep isohalines; the slope is $5 \times 10^{-3}$. The plume remains in contact with the bottom. The simulated coastal current (Fig. 7f) possesses a larger salinity anomaly than observed along the cross-section. Simulation results exhibit the same isohaline slope seen in observations. The degree of bottom contact is extensive ($Y_s/Y_r$ is 0.8) and is 14% greater than observed. Though the simulation produces a shorter buoyant outflow than observed at this time, the salinity range, width, and amount of bottom contact are comparable to observations. Simulation results exhibit the observed response to the downwelling wind event.

### c. DCC2b mapping: High river discharge, upwelling winds

During the DCC2b mapping (1–2 May 1993), the buoyant outflow responded to moderate upwelling-favorable winds. The high river discharge conditions of the previous mapping persisted. The switch to up-

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**Table 3.** Downshelf extent and degree of bottom contact. Downshelf extent of the 31-psu ($X_{31}$) and 29-psu ($X_{29}$) isohalines are derived from the Fig. 6 surface salinity fields. Bottom contact is measured by comparing offshore distances of the 31-psu isohaline at the surface ($Y_s$) and bottom ($Y_r$) for the Fig. 7 cross section.

<table>
<thead>
<tr>
<th>Mapping (1993)</th>
<th>$X_{31}$ (km)</th>
<th>$X_{29}$ (km)</th>
<th>$Y_s/Y_r$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Observed</td>
<td>Model</td>
<td>Observed</td>
</tr>
<tr>
<td>DCC1</td>
<td>190</td>
<td>110</td>
<td>0.7</td>
</tr>
<tr>
<td>DCC2a</td>
<td>125</td>
<td>75</td>
<td>0.7</td>
</tr>
<tr>
<td>DCC3</td>
<td>130</td>
<td>105</td>
<td>0.5</td>
</tr>
<tr>
<td>DCC4</td>
<td>$&gt;95$</td>
<td>75</td>
<td>$&lt;0.2$</td>
</tr>
</tbody>
</table>
wellington winds (peaking at 4 m s\(^{-1}\)) reduced downshelf transport and widened the plume (Fig. 6c), but was not strong enough to dominate over discharge influence and reverse the coastal current.

The maximum salinity anomaly (8.2 psu) is stronger than seen in either the DCC1 or DCC2a mappings. The plume has widened in response to offshore surface Ekman transport (\(K\) is between 1.7 and 5.7). Weak wind-driven upshelf flow has held the 31-psu isohaline nearly in place, while the coastal current continued to push more buoyant waters downshelf. The 29-psu isohaline has penetrated 30 km farther downshelf since the DCC2a mapping. The buoyancy signatures of the observed and simulated plume (Fig. 6g) are similar; maximum salinity anomalies agree within 0.1 psu (1%). Plume width in the model (between 9 and 27 km) is wider than simulation results during DCC2a, revealing that the model captures offshore spreading in response to moderate upwelling-favorable winds. Plume width (\(K\) ranges from 1.5 to 4.5) is similar to observed values except near mouth. The downshelf extent of the 31- and 29-psu isohalines is only 20% shorter than observed.

The observed vertical salinity structure (Fig. 7c) has higher stratification and smaller isohaline slopes (between \(1 \times 10^{-3}\) and \(2 \times 10^{-3}\)) than in DCC2a. Upwelling winds also have decreased bottom contact to intermediate levels (\(Y_s/Y_f\) is 0.5). The simulated coastal current (Fig. 7g) exhibits the observed pattern from DCC2a to DCC2b in response to upwelling winds: a wider plume, increased vertical stratification, decreased isohaline slope, and less bottom contact. Simulation results possess a characteristic isohaline slope of \(1 \times 10^{-3}\) and intermediate bottom contact (\(Y_s/Y_f\) is 0.6). While there is some mismatch between observations and simulation results, it is encouraging that the model produces a buoyant outflow with similar dimensions, buoyancy signature, bottom contact, and sensitivity to winds.

### Table 4. Statistical evaluation of model performance with respect to surface salinity.

<table>
<thead>
<tr>
<th>Mapping</th>
<th>(\sigma_{\text{obs}}) (psu)</th>
<th>(\sigma_{\text{model}}) (psu)</th>
<th>Bias/(\sigma_{\text{obs}})</th>
<th>Rmse/(\sigma_{\text{obs}})</th>
</tr>
</thead>
<tbody>
<tr>
<td>DCC1</td>
<td>1.3</td>
<td>1.6</td>
<td>0.45</td>
<td>0.88</td>
</tr>
<tr>
<td>DCC2a</td>
<td>0.9</td>
<td>1.0</td>
<td>0.22</td>
<td>0.73</td>
</tr>
<tr>
<td>DCC2b</td>
<td>1.6</td>
<td>1.9</td>
<td>0.05</td>
<td>0.76</td>
</tr>
<tr>
<td>DCC3</td>
<td>0.5</td>
<td>0.6</td>
<td>0.89</td>
<td>1.19</td>
</tr>
<tr>
<td>DCC4</td>
<td>1.3</td>
<td>2.2</td>
<td>0.01</td>
<td>1.06</td>
</tr>
<tr>
<td>DCC5</td>
<td>3.9</td>
<td>3.2</td>
<td>0.46</td>
<td>0.58</td>
</tr>
<tr>
<td>DCC6</td>
<td>1.5</td>
<td>1.6</td>
<td>-0.15</td>
<td>0.72</td>
</tr>
<tr>
<td><strong>All</strong></td>
<td><strong>2.0</strong></td>
<td><strong>1.9</strong></td>
<td><strong>0.21</strong></td>
<td><strong>0.62</strong></td>
</tr>
</tbody>
</table>

The maximum salinity anomaly of the observed surface salinity field (Fig. 6d) is 3.5 psu. Plume width is greater than 60 km (beyond the sampled region); the Kelvin number exceeds 10. The 30-psu isohaline extends 40 km offshore and 50 km downshelf. The plume differs greatly from the slender outflows observed during the other 1993 surveys. The observed plume and simulated outflow (Fig. 6h) exhibit similar buoyancy signatures and dimensions. The maximum salinity anomaly is within 10% of the observed value. As in observations, the 31-psu contour is more than 60 km offshore. The 30-psu isohaline extends 30 km offshore and 40 km downshelf. Simulation results reveal that buoyant waters have been transported upshelf and offshore due to upwelling winds.

Observed vertical structure during these low discharge conditions is shown in Fig. 7d. Upwelling winds prior to the survey have spread buoyant waters offshore in a 10-m-thick mixed layer. Isohalines are nearly horizontal beneath much of the plume. Bottom contact occurs out to 18-m depth; however, most of the buoyant outflow does not contact the bottom (\(Y_s/Y_f < 0.2\)). Simulation results (Fig. 7h) are similar to observations: the simulation produces a wide plume with horizontal isohalines and limited bottom contact (\(Y_s/Y_f < 0.2\)).

e. **Statistical comparison**

Model performance can be evaluated statistically through comparison to the 1993/94 shelf salinity surveys. Following Oke et al. (2002), the standard deviation of the observations (\(\sigma_{\text{obs}}\)) is compared with the standard deviation of corresponding model results (\(\sigma_{\text{model}}\)). Simulation mismatch is characterized by model bias (\(\bar{S}_{\text{model}} - \bar{S}_{\text{obs}}\)) and root-mean-square error (rmse). As in Oke et al. (2002), bias and error values are non-dimensionalized by \(\sigma_{\text{obs}}\) to generalize results. Table 4 reports statistical results for surface salinities along survey transects. Table 5 gives the statistical comparison of salinity profiles at CTD locations.
Standard deviations of observations and simulation results are similar for each mapping (Tables 4 and 5): the $\sigma_{\text{obs}}$ range is 0.5–3.9 psu and $\sigma_{\text{model}}$ is between 0.5 and 3.2 psu. Standard deviations for the entire dataset agree within 0.1 psu (5% of $\sigma_{\text{obs}}$). There is a positive model bias in surface salinities: 0.4 psu (21% of $\sigma_{\text{obs}}$). This bias is less than 10% of the typical plume buoyancy signature (5 psu). Even though the simulated plume is fresher than observed, it tends to be shorter and the region outside the active plume (most of the sampled area) is slightly saltier than observed. Model bias is smaller for the salinity profiles: 0.03 psu (2% of $\sigma_{\text{obs}}$). The rms errors for the surface salinity and salinity profile data are 1.2 psu (62% of $\sigma_{\text{obs}}$) and 0.9 psu (67% of $\sigma_{\text{obs}}$), respectively. Model bias and rms errors are comparable to levels found by Oke et al. (2002) in simulations of upwelling circulation along the Oregon coast.

Linear regression results between observed and simulated surface salinities (for the seven 1993/94 mappings) indicate a high correlation of $r = 0.82 (0.69 < r < 0.90; N_{\text{eff}} = 40)$; the regression slope is $0.79 \pm 0.18$. Comparing salinity profiles gives similar results; $r$ is $0.78 (0.70 < r < 0.81; N_{\text{eff}} = 251)$ and $b$ is $0.79 \pm 0.08$. Observed and simulated plume characteristics such as buoyancy signature, stratification, width, and downshelf extent also can be compared via linear regressions. Figure 8 includes the correlation coefficient and regression slope for each of these derived variables; $r$ and $b$ values are shown with confidence intervals based on $N_{\text{eff}}$. Buoyancy signature is represented by the maximum surface salinity along each transect ($\Delta S_{\text{max}}$). Simulated and observed $\Delta S_{\text{max}}$ are highly correlated; $r$ is $0.85$. The regression slope is $0.88 \pm 0.30$ (Fig. 8a). Although $b$ is near 1, the simulation tends to be fresher than observed at higher $\Delta S_{\text{max}}$ and not as fresh at lower $\Delta S_{\text{max}}$ because the simulated plume typically is fresher and shorter than observed. Stratification is assessed with the vertical salinity range of each CTD profile ($\Delta S_{\text{vert}}$). The correlation coefficient for $\Delta S_{\text{vert}}$ is 0.68 and the regression slope is $0.77 \pm 0.31$ (Fig. 8b). Plume width is determined by the offshore distance of the 31-psu isohaline at each transect (only nonzero widths are included in the regression). There is good correlation between the simulated and observed widths; $r$ is 0.77. The regression slope is $0.83 \pm 0.33$ (Fig. 8c), indicating the simulated plume tends to be only 17% narrower than observed. Downshelf extent of isohalines (26–31 psu at a 1-psu interval) is calculated from contoured surface salinity fields. Simulated and observed downshelf extents are correlated with $r = 0.85$, but the simulated plume tends to be shorter than observed ($b = 0.63 \pm 0.38$).

### Table 5. Statistical evaluation of model performance with respect to salinity profiles.

<table>
<thead>
<tr>
<th>Mapping</th>
<th>$\sigma_{\text{obs}}$ (psu)</th>
<th>$\sigma_{\text{model}}$ (psu)</th>
<th>Bias/(\sigma_{\text{obs}})</th>
<th>Rmse/(\sigma_{\text{obs}})</th>
</tr>
</thead>
<tbody>
<tr>
<td>DCC1</td>
<td>1.4</td>
<td>1.5</td>
<td>0.36</td>
<td>0.80</td>
</tr>
<tr>
<td>DCC2a</td>
<td>0.6</td>
<td>0.5</td>
<td>-0.18</td>
<td>0.74</td>
</tr>
<tr>
<td>DCC2b</td>
<td>1.0</td>
<td>1.1</td>
<td>-0.20</td>
<td>0.71</td>
</tr>
<tr>
<td>DCC3</td>
<td>0.8</td>
<td>0.7</td>
<td>0.01</td>
<td>0.50</td>
</tr>
<tr>
<td>DCC4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>DCC5</td>
<td>2.6</td>
<td>2.0</td>
<td>0.06</td>
<td>0.63</td>
</tr>
<tr>
<td>DCC6</td>
<td>1.4</td>
<td>1.5</td>
<td>-0.29</td>
<td>0.72</td>
</tr>
<tr>
<td>All</td>
<td>1.4</td>
<td>1.3</td>
<td>0.02</td>
<td>0.67</td>
</tr>
</tbody>
</table>

9. **Lagrangian observations of the coastal current**

In this section, simulated drifter trajectories are compared with data from four 1993 drifter releases (times shown in Figs. 2 and 5). Satellite-tracked surface drifters (drogued at 0.5-m depth) were deployed outside the Delaware Bay mouth to sample the coastal current system during a range of river discharge and wind conditions. Simulated drifters were initialized at the same time, location, and depth as the actual drifters. Velocity fields used to advect drifters were from hourly model results at the drifter drogue depth. Simulated drifter positions were updated every 15 minutes. Comparisons focus on the first three days after each release in order to easily characterize wind conditions and describe trajectories. Figure 9 graphs trajectories and along-path salinities. The comparisons in this section concentrate on how well the simulation reproduces the drifter cloud, rather than on individual trajectories.

The first drifter release (25 March 1993) occurred prior to the spring high discharge period; lagged river discharge was below 1000 m$^3$ s$^{-1}$ (Fig. 5). Downwelling winds peaking at 7 m s$^{-1}$ blew during this period (Fig. 2b). All drifters move downshelf after release (Fig. 9a). After initially following the ancestral channel offshore, many of the drifters grounded near 25 km downshelf. The onshore transport evidently is due to surface Ekman transport induced by downwelling winds. These winds also drive downshelf flow that advects the remaining drifters farther downshelf. Lagrangian salinities are high (28.0–30.5 psu) because drifters left the source region before much freshwater reached the shelf; they were released prior to the high discharge conditions characterizing the DCC1 mapping (section 8a). The corresponding simulated drifter results are shown in Fig. 9e. Most drifters follow observed paths and ground near the same location downshelf. Only one drifter escapes grounding; it progresses farther downshelf under the influence of wind-driven down-
shelf flow. The limited grid resolution likely prevented more drifters from avoiding grounding in the face of onshore surface Ekman transport. The simulation exhibits the observed downshelf and onshore transport of the drifters, but salinities are uniformly higher (between 29.6 and 31.1 psu). The observed and simulated DCC1 salinity fields share this difference downshelf of the strong buoyant outflow. This indicates that, prior to high river discharge rates, buoyant waters in the simulation are more mixed than observed.

The second drifter release took place on 23 April 1993 (a week prior to DCC2a) under high river discharge conditions (between 1700 and 2800 m³ s⁻¹) and initially light winds. During the deployment, a strong upwelling wind event (11 m s⁻¹ peak winds) occurred. The drifter data (Fig. 9b) highlight wind influence on the Delaware coastal current. Initially, drifters move downshelf in the buoyancy-driven current. Building upwelling winds arrest this downshelf transport. Winds are powerful enough to reverse the entire coastal current and advect drifters upshelf and offshore. After this reversal, salinities increase along drifter paths from 26.7 up to 32.0 psu. This points to mixing during the upwelling event (Sanders 1999).

Simulated drifters (Fig. 9f) first progress downshelf. They then travel upshelf and offshore, exhibiting the

---

**Fig. 8. Regression analyses.** (a) Maximum salinity anomaly along each survey transect. (b) Vertical salinity range for each salinity profile. (c) Plume width (following the 31-psu isohaline) at each transect (only nonzero data points are included). (d) Downshelf extent of isohalines (from 26 to 31 psu at a 1-psu interval). The correlation coefficient \( r \) (with the 95% confidence interval in parentheses), the regression slope \( b \) with its 95% confidence interval, and the number of data points \( N \) (with \( N_{eff} \) in parentheses) are listed in each graph. The regression line (solid) and one-to-one fit line (dashed) also are shown. Observations and corresponding simulation results are from the seven 1993/94 shelf salinity mappings.
observed reversal and increasing salinities in response to the upwelling winds. The simulated drifters follow the observed trajectories (though less dispersed) but move farther upshelf. The simulated and observed drifter paths are evidence of wind-induced reversals of the coastal current during high discharge conditions.

The third drifter deployment (10 May 1993) coincided with moderate river discharge (900 m³ s⁻¹) and upwelling-favorable winds (peaking at 6 m s⁻¹). Drifters are advected offshore with little alongshelf dispersion (Fig. 9c). As in the previous release, salinities increase along drifter paths (rising from 26.7 to 32.0 psu). The corresponding simulated drifter release (Fig. 9g) resembles observations in terms of offshore wind-driven transport. Alongshelf drifter dispersion is larger than observed. Salinity increases along drifter paths as observed; however, the buoyancy signature is maintained farther offshore in the model.

The final 1993 deployment began on 27 May 1993 (several days before the DCC3 mapping). River discharge (below 300 m³ s⁻¹) was low and winds were upwelling-favorable (5 m s⁻¹) after release. This release (Fig. 9d) samples a weak buoyant outflow under the influence of moderate upwelling winds. Drifters initially move along the ancestral channel and downshelf because the release coincided with light winds. As winds build, the drifters move offshore and turn upshelf. Salinities increase along drifter paths from 27.0 up to 30.9 psu. The simulated drifter field (Fig. 9h) exhibits the initial downshelf transport followed by wind-driven offshore advection. The along-path increase in salinity (27.0–30.2 psu) is consistent with the observed salinity pattern.

10. Summary and conclusions

This study evaluates simulation performance in light of estuarine and shelf observations. Observational data used for comparison include an estuarine salinity climatology, a freshwater flux time series recorded near the bay mouth, shelf salinity mappings, CTD profiles, and surface drifter data.

Comparisons with estuarine salinity data indicate conditions at the mouth and in the lower bay are consistent with observations but the along-estuary salinity gradient is underestimated. The simulation exhibits the observed lateral salinity pattern produced by gravitational circulation interacting with bay bathymetry. Es-
tuarine salinity results possess the observed vertical structure and weak response to river discharge. Thus, the simulation performs well enough to provide representative source conditions for the Delaware buoyant outflow.

Field observations of the Delaware coastal current collected during spring 1993 and 1994 afford a rare opportunity to rigorously evaluate model performance. Current and salinity data recorded along the mooring arc (positioned outside the bay mouth) provide a measurement of freshwater delivery to the shelf. The simulated and observed time series of freshwater flux are highly correlated and exhibit the same amplitude of response to river discharge and wind forcing. Comparisons with shelf surface salinities reveal that the simulated and observed plumes have comparable buoyancy signatures and dimensions for a variety of river discharge and wind conditions. The most pronounced difference is that the model tends to produce a shorter and fresher plume. Simulation results are consistent with the observed variability in vertical structure: isopycnal slope and degree of bottom contact are similar for each mapping. Simulated drifter deployments exhibit the observed onshore transport because of downwelling-favorable winds. Simulated drifters also follow the observed offshore and upshelf tracks while under the influence of strong upwelling-favorable winds. Drifter dispersion, however, typically is lower in the simulation; increased grid resolution may alleviate this mismatch.

Shelf salinity mappings provide enough data to statistically evaluate model performance. Observed and simulated salinities are highly correlated ($r = 0.78$ for surface values and $r = 0.82$ for vertical profiles) with a regression slope of 0.79 ($\pm 0.18$ for surface values and $\pm 0.08$ for profiles). The standard deviations of observed and simulated surface salinity fields and salinity profiles agree within 0.1 psu. Model bias is less than 10% of the typical plume buoyancy signature. Rms error (for all mappings) is 1.2 psu for surface salinities and 0.9 psu for profile data; this error indicates a point-to-point mismatch in the salinity data. This mismatch arises from differences in plume buoyancy signature and dimensions. As is apparent in Fig. 8a, the simulation tends to overestimate buoyancy in the freshest plume waters (closer to the mouth) and underestimate salinity anomalies in saltier waters (farther downshelf). Simulated and observed maximum salinity anomaly, vertical salinity range, plume width, and downshelf extent are well correlated: $r$ is 0.85, 0.68, 0.77, and 0.85, respectively. Regression slopes for these plume characteristics are $0.88 \pm 0.30$, $0.77 \pm 0.31$, $0.83 \pm 0.33$, and $0.63 \pm 0.38$. These results indicate the simulation somewhat underestimates the observed variability in buoyancy signature and plume dimensions.

The simulated plume exhibits observed patterns in response to different discharge levels and wind conditions. A slender coastal current propagates downshelf under high discharge conditions and light or downwelling-favorable winds. Low discharge conditions create a weak buoyant outflow that is highly susceptible to winds. Downwelling-favorable winds act to compress the plume against the coast, steepen isopycnals, and increase bottom contact. Upwelling winds act to spread buoyant waters offshore, flatten isopycnals, and decrease bottom contact. Strong enough upwelling winds can reverse the coastal current. Despite significant differences between the simulated and mapped plume, in each case the simulation produces a buoyant outflow with a buoyancy signature, width, and extent similar to the observed coastal current. As in observations, isohalines are steeper during downwelling-favorable winds and more gently sloped during upwelling-favorable winds.

Simulation results are consistent with the observed dynamics of the Delaware buoyant outflow. This is in contrast with the performance of an earlier simulation attempt (using an early version of POM) by Galperin and Mellor (1990a,b,c). Münchow and Garvine (1993a) note that these “model results clash with our observational evidence” of the Delaware coastal current. Key factors leading to the poor model performance are surface elevation and mass flux constraints imposed across the bay mouth (inside the domain) to match sea level observations.

Since the present simulation is generally consistent with field observations, it should be a useful tool for investigating the dynamics of river discharge, wind, and tidal forcing of the Delaware buoyant outflow. This is the focus of Whitney (2003) and forthcoming publications, including Whitney and Garvine (2005).

The model application would benefit from removal of several present limitations. First, the model neglects spatial variation in the large-scale wind-driven shelf circulation. Alongshelf pressure gradients associated with synoptic weather systems and remotely forced shelf waves impact wind-driven circulation in the Middle Atlantic Bight (Noble and Butman 1983; Yankovsky and Garvine 1998; Wong 1999) but are omitted in the present simulation. Second, the present grid resolution cannot be used to study fronts since across-front length scales typically are of 10-m order (Marmorino and Trump 2000). Consequently, model performance with respect to these frontal features has not been assessed. Third, the model neglects ambient shelf stratification, which grows important during summer months.
Nevertheless, this model evaluation indicates that ECOM3d and POM can simulate a wide range of observed coastal current features and are a competent mechanism for revealing dynamics of coastal buoyant outflows. Systematic disagreement between simulation results and observations is present: the simulated plume tends to be fresher and shorter than observed, indicating less mixing in the model. This may indicate generic shortcomings in the Mellor–Yamada mixing scheme, such as its premature shutoff at high Richardson numbers (Garvine 1999; Canuto et al. 2001). Mixing in buoyant plumes is a subject of ongoing modeling and observational research.

This study helps to fill a conspicuous void by performing a systematic evaluation of a standard numerical model in a coastal buoyant outflow application. The model assessment provides perspective for interpreting results and establishes the simulation as a useful tool in studying the Delaware coastal current. This application provides some confidence that ECOM3d and POM can usefully simulate coastal currents generically; however, it does not validate the model for other coastal buoyant outflows. Rigorous comparisons between observations and simulation results for other coastal currents are encouraged.

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