

Physical processes controlling the formation, evolution, and perturbation of the low-salinity front in the inner shelf off the southeastern United States: A modeling study

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Abstract. Physical processes that control the formation, evolution, and perturbation of the low-salinity front over the inner shelf of the South Atlantic Bight have been examined using a fully three-dimensional primitive equation and turbulent closure model. The model was forced by semidiurnal tides (M_2 , S_2 , and N_2), climatological means of multiple river discharges, and upwelling-favorable wind. This model has provided a reasonable simulation of the fortnightly and monthly variations of semidiurnal tides. Computed amplitudes and phases of tides show good agreement with observational data available at tidal gauge stations along the coast. Spatial structures of buoyancy currents are significantly modified by stratified tidal rectification. As a result, the southward residual current intensifies significantly at the front and reduces or reverses close to the coast. A weak velocity area forms in the downstream region of each river, which blocks the low-salinity water to form multiple "tongue-like" domes in the inner shelf. For a given springtime climatological upwelling-favorable wind, isolated low-salinity lenses can form episodically in two steps. At first, a geometrically controlling wave-like frontal shape develops at the outer edge of the frontal zone as a result of the interaction between tides, multiple river discharges, and upwelling-favorable wind. Then, isolated low-salinity lenses form at the crest when water on the shoreward side of the crest is displaced by relatively high salinity water advected from the upstream trough south of the crest and diffused upward from the deeper region. Wind-induced upwelling is noticeable to compensate for the water loss due to the near-surface offshore Ekman transport, but it does not play a critical role in the formation of isolated lenses under the climatological conditions of river discharges and upwelling-favorable winds over the inner shelf of the South Atlantic Bight.

1. Introduction

The South Atlantic Bight (SAB) refers to the continental shelf bounded to the north at 35°N, Cape Hatteras, North Carolina and to the south at 27°N, West Palm Beach, Florida. It is a typical "bent" shelf with a width of about 5 km off Palm Beach, 120 km off Georgia and South Carolina, and 30 km off Cape Hatteras [Menzel, 1993]. According to physical processes that control the water properties, the SAB can be divided into three oceanographically climatological zones: in-

ner, middle, and outer shelves [Atkinson and Menzel, 1985]. The inner shelf, bounding at the 20-m isobath, is mainly characterized by a coastal low-salinity frontal zone, which results from the interaction between freshwater discharges, tidal mixing, and wind forcing. The outer shelf, however, is dominated by the shelf break front between the Gulf Stream and coastal water [Lee and Atkinson, 1983]. The midshelf, located in the range of the 20- to 40-m isobath, is a zone that is controlled by the combined processes on inner and outer shelves (Figure 1).

Ten major rivers terminate along the coast of the SAB (Figure 1). Although the freshwater discharge from any one of these rivers is smaller compared with that of the Mississippi River over the Louisiana and Texas shelf [Dinnel and Wiseman, 1986], the total fresh-

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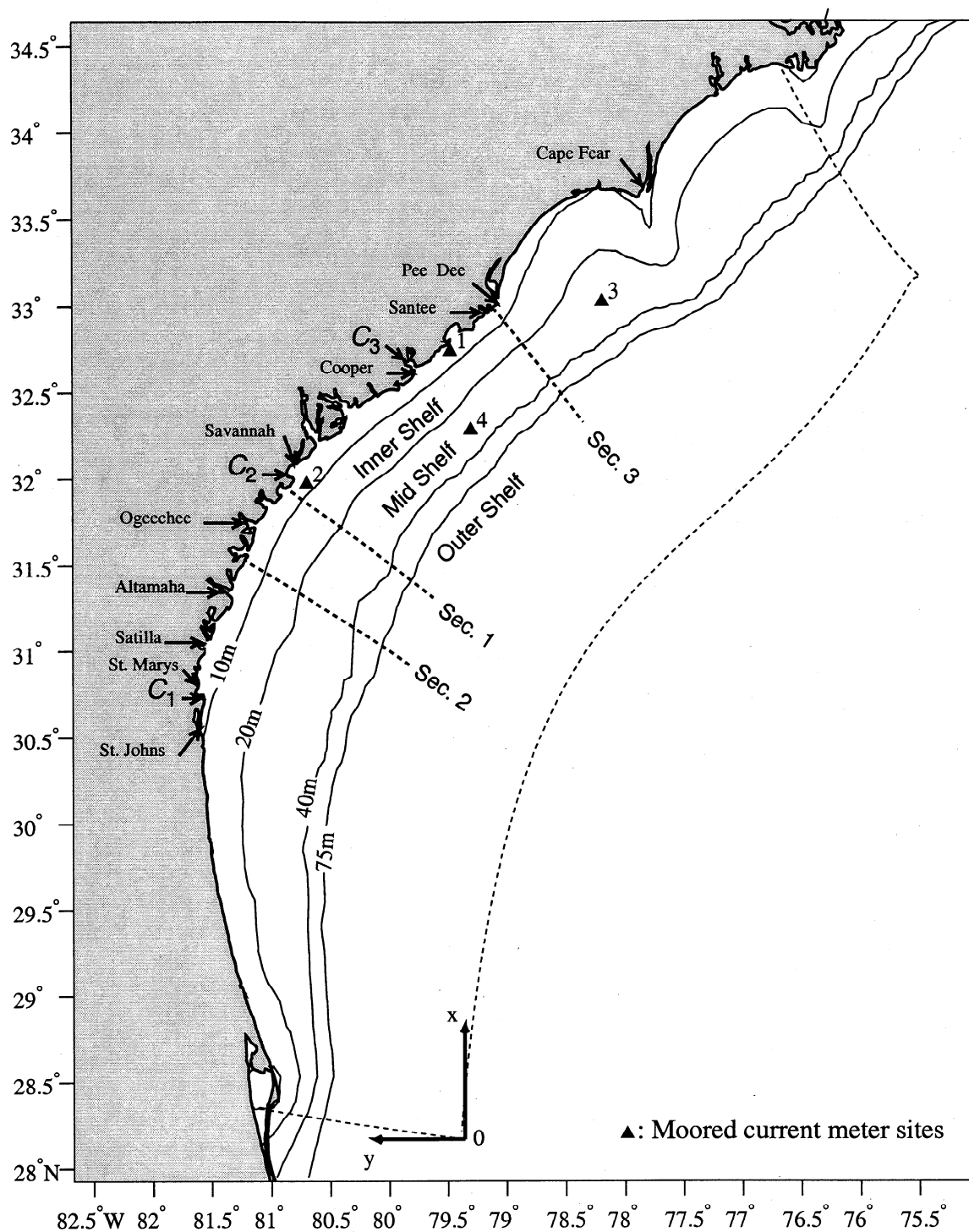


Figure 1. Continental shelf off the southeastern United States and locations of the 10 rivers. The dashed line indicates the domain of the inner shelf model. Two cross-shelf sections are represented by thick lines. C_1 , C_2 , and C_3 are the tidal stations at Fernandina Beach, Florida, Savannah, Georgia, and Charleston, South Carolina. The numerical domain consists of 145 (along shelf) $\times$ 100 (cross shelf) $\times$ 16 (vertical) grid points, with horizontal resolutions of 2 to 3 km in the cross-shelf direction and 4 to 6 km in the along-shelf direction. Solid triangles indicate the mooring stations used for model and data comparisons in this study.

water discharge is comparable. The 20-year mean of total discharge shows a seasonal low of 1000 m³/s in autumn and a maximum of 4000 m³/s in spring [Blanton, 1981]. The maximum discharge reached more than

7000 m³/s during a high discharge year like 1993 [Blanton, 1996].

The inner shelf of the SAB is dominated by tidal motion. For example, M_2 tidal currents are about 30 to

40 cm/s near the coast. These currents account for about 80 to 90% (cross shelf) and about 20 to 40% (along shelf) of current variation and kinetic energy in the inner shelf [Tebeau and Lee, 1979; Lee and Brook, 1979; Pietrafesa et al., 1985]. Tidal energy is largest in the widest part of the shelf between Savannah and Charleston and smallest at the northern and southern ends.

Tidal mixing is energetic in the inner shelf of the SAB, which significantly contributes to vertical mixing within the low-salinity plume, and causes a surface-bottom front in the inner SAB shelf. In addition, tidal rectification over the variable bottom shelf tends to produce the residual flow over the inner shelf [Huthnance, 1973; Zimmerman, 1978; Loder, 1980]. The residual current can be intensified at the front by stratified tidal rectification through the nonlinear interaction between barotropic and internal tides, reduced internal friction, and an enhanced cross-frontal density gradient [Chen and Beardsley, 1995]. The dynamic processes of the baroclinic response of the SAB to stratified tidal rectification, however, are still not well understood.

The climatology of wind fields over the SAB can be divided into five seasons: winter (November to February), spring (March to May), summer (June to July), fall (August), and mariners' fall (September to October). The entire shelf of the SAB is characterized by northwesterly, northerly, or northeasterly wind in winter and by southeasterly or southwesterly wind in summer. Spring and fall seasons are in the transitional regime, which is typified by a relatively weak wind except during atmospheric frontal passages. In mariners' fall, the northeasterly wind, which is strongest for the entire year [Weber and Blanton, 1980], prevails over the entire SAB.

Past hydrographic measurements showed that some relatively fresh lenses were formed at the outer edge of the low-salinity front in the inner shelf of the SAB in spring, when the upwelling-favorable wind prevailed over the shelf [Blanton, 1996; Blanton et al., 1989; Nelson and Guarda, 1995]. These lenses tended to be isolated or detached from the front and extended offshore to the midshelf. This phenomenon was detected in the modeling simulation of the 1984 field data in this region by Kourafalou et al. [1996]. Their works suggest that the possible offshore detachment of isolated lenses from the frontal zone appears to account for a large portion of the cross-frontal water transport over the inner SAB shelf in spring. The physical mechanisms responsible for the formation of these isolated lenses, however, have not been well explored.

In this paper, we examine the physical processes that control the formation, evolution, and perturbation of the low-salinity front in the inner shelf of the SAB. Numerical experiments were conducted over a three-dimensional (3-D) real geometric shelf that did not include the Gulf Stream at the shelf break. It is helpful to study an isolated inner shelf system in order to under-

stand the dynamics of the complex nonlinear interactions between tide, wind, and multiple river discharges over the inner shelf of the SAB.

The remaining sections of this paper are organized as follows. Section 2 describes the 3-D numerical model and the numerical experiment designs. Section 3 represents the semidiurnal tidal simulation and barotropic tidal rectification. In section 4, we describe the formation of the low-salinity front in the inner shelf for cases with and without tides. In section 5, the cross-frontal water exchanges are examined for an upwelling-favorable wind, and conclusions are summarized in section 6.

2. Design of Numerical Experiments

The numerical model used in this study is an updated modified version of the 3-D coastal ocean circulation model originally developed by Blumberg and Mellor [1987]. This is a fully nonlinear, prognostic model incorporating the free surface and realistic parameterization of vertical mixing through the Mellor and Yamada [1974, 1982] level 2.5 turbulent closure scheme (so-called MY level 2.5 model). A modification of the stability functions made by Galperin et al. [1988] is included in this updated version. The numerical code of this model is described in detail by Blumberg [1994], and only a brief model formula is given here.

The model consists of momentum, continuity, temperature, salinity, and density equations:

$$\frac{\partial u}{\partial t} + \mathbf{v} \cdot \nabla u - f v = -g \frac{\partial \zeta}{\partial x} - \frac{\partial P}{\partial x} + \frac{\partial}{\partial z} \left(K_m \frac{\partial u}{\partial z} \right) + F_u \quad (1)$$

$$\frac{\partial v}{\partial t} + \mathbf{v} \cdot \nabla v + f u = -g \frac{\partial \zeta}{\partial y} - \frac{\partial P}{\partial y} + \frac{\partial}{\partial z} \left(K_m \frac{\partial v}{\partial z} \right) + F_v \quad (2)$$

$$\frac{\partial \zeta}{\partial t} + \frac{\partial [u(\zeta + H)]}{\partial x} + \frac{\partial [v(\zeta + H)]}{\partial y} = 0 \quad (3)$$

$$\frac{\partial \theta}{\partial t} + \mathbf{v} \cdot \nabla \theta = \frac{\partial}{\partial z} \left(K_h \frac{\partial \theta}{\partial z} \right) + F_\theta \quad (4)$$

$$\frac{\partial s}{\partial t} + \mathbf{v} \cdot \nabla s = \frac{\partial}{\partial z} \left(K_h \frac{\partial s}{\partial z} \right) + F_s \quad (5)$$

$$\rho_{\text{total}} = \rho_{\text{total}}(\theta, s) \quad (6)$$

where x , y , and z are along-shelf, cross-shelf, and vertical axes of the Cartesian coordinate (Figure 1); u , v , and w are x , y , z velocity components; $\mathbf{v} = u\mathbf{i} + v\mathbf{j} + w\mathbf{k}$; ∇ is the gradient operator; θ is potential temperature; s is salinity; ζ is sea surface elevation; P is baroclinic pressure defined as $1/\rho_0 \int_z^0 \rho g dz$; f is Coriolis parameter; g is the gravitational acceleration; K_m is the vertical eddy viscosity coefficient; and K_h is the thermal vertical eddy friction coefficient. F_u , F_v , F_θ and F_s represent the horizontal momentum, thermal, and salt diffusion terms, while ρ and ρ_0 are the perturbation and reference density, which satisfy

$$\rho_{\text{total}} = \rho + \rho_o \quad (7)$$

K_m and K_h are calculated using the MY level 2.5 turbulent closure scheme. This turbulent closure model is designed to simulate boundary layer physics through the inclusion of (1) shear and buoyancy production of turbulent kinetic energy, (2) dissipation of turbulent kinetic energy, (3) vertical diffusion and the time derivative of the turbulent kinetic energy and turbulent macroscale, and (4) the advection of the turbulent kinetic energy. F_u , F_v , F_θ and F_s are calculated by Smagorinsky's [1963] formula, which is directly proportional to the product of horizontal grid sizes. See Blumberg and Mellor [1987] and Chen and Beardsley [1995] for a detailed description of the MY level 2.5 turbulent closure scheme.

In the absence of surface and bottom heat fluxes, the surface and bottom boundary conditions for the momentum, heat, and salt equations are

$$K_m \left(\frac{\partial u}{\partial z}, \frac{\partial v}{\partial z} \right) = \frac{1}{\rho_o} (\tau_{sx}, \tau_{sy}), \quad \frac{\partial \theta}{\partial z} = \frac{\partial s}{\partial z} = 0, \\ w = \frac{\partial \zeta}{\partial t} + u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y}, \quad z = \zeta(x, y, t), \quad (8)$$

$$K_m \left(\frac{\partial u}{\partial z}, \frac{\partial v}{\partial z} \right) = \frac{1}{\rho_o} (\tau_{bx}, \tau_{by}), \quad \frac{\partial \theta}{\partial z} = \frac{\partial s}{\partial z} = 0, \\ w = -u \frac{\partial H}{\partial x} - v \frac{\partial H}{\partial y}, \quad z = -H(x, y) \quad (9)$$

where (τ_{sx}, τ_{sy}) and $(\tau_{bx}, \tau_{by}) = C_d \sqrt{u^2 + v^2}(u, v)$ are the x and y components of surface wind and bottom stresses. The surface wind stress is calculated based on the neutral, steady state drag coefficient developed by Large and Pond [1981]. The drag coefficient C_d is determined by matching a logarithmic bottom layer to the model at a height z_{ab} above the bottom, i.e.,

$$C_d = \max \left[k^2 / \ln \left(\frac{z_{ab}}{z_o} \right)^2, 0.0025 \right]$$

where $k = 0.4$ is von Karman's constant and z_o is the bottom roughness parameter, taken here as $z_o = 0.1$ cm.

The model domain is shown in Figure 1, which is bounded to the north at Cape Hatteras, North Carolina and to the south at Cape Canaveral, Florida. The offshore open boundary is specified using a curvature line parallel with the coastal boundary and with a maximum depth cutoff of 200 m. The horizontal model grids are designed using an orthogonal curvilinear coordinate transformation with total grid points of 145 (along shelf) $\times$ 100 (cross shelf), which provides a horizontal resolution of about 2 to 3 km in the cross-shelf direction and 4 to 6 km in the along-shelf direction. The total grid number of points in the vertical is 16. This results in a vertical resolution of about 0.2 m near the coast and 13 m at the 200-m isobath. A σ coordinate transformation defined as $\sigma = (z - \zeta)/(H + \zeta)$ is used in numerical experiments.

The model was forced initially by semidiurnal tidal elevations (M_2 , S_2 , and N_2) at three open boundaries. The sea level data used for tidal forcing were directly adopted from the global $0.5^\circ \times 0.5^\circ$ tidal model developed recently by Egbert *et al.* [1994]. To simplify the model problem and focus on physical processes associated with the low-salinity front, we ignored the spatial structure of the background temperature by using a constant value of 20°C . We also ignored the spatial variation of the background salinity initially by using a constant value of 35 practical salinity units (psu). This simplification is consistent with our assumption that any spatial variation of water salinity in the inner shelf should develop as a result of multiple river discharges and tidal and wind mixings.

After tidal currents reached a quasi-equilibrium state, the fresh waters were injected into the model domain at the coastal boundaries of 10 rivers as volume fluxes defined as

$$V_0(t) = -\Delta x \int_{-H_0}^{\zeta} v_0 dz \quad (10)$$

where Δx is the along-shelf grid resolution and v_0 , ζ , and H_0 are the cross-shelf current, the sea surface elevation, and the mean water depth at point sources, respectively. The discharge rates of 10 rivers were taken from the annual means of stream gauge data recorded over a period of 7 to 74 years (Table 1). The temperature of fresh water was the same as that of the background temperature used over the shelf.

We also examined the effects of upwelling-favorable wind on the temporal and spatial variations of the low-salinity front for the cases with and without tides. A constant southwesterly wind with a magnitude of 5 m/s was added after the low-salinity front was fully developed in the inner shelf. The model initially was ramped up over a M_2 tidal cycle to avoid the sharp transient associated with the initial forcing conditions [Chen and Beardsley, 1995]. The time step used in our numerical experiments was 931.5 s.

3. Simulation of Semidiurnal Tides and Residual Currents

To simulate the fortnightly and monthly variations of semidiurnal tides, we ran the model with three semidiurnal tidal forces (M_2 , S_2 , and N_2) for over 100 model days. The model-predicted time series of surface elevation and currents at each grid point was fitted by a least squares harmonic analysis method. The resulting coamplitude and cophase of each tidal constituent are shown in Figure 2.

Tides propagate toward the coast from the open ocean and turn clockwise as they reach the inner shelf. The M_2 tide reaches a maximum amplitude of about 100 cm near the coast in the midshelf between Georgia and South Carolina and decreases offshore and toward both south and north ends of the shelf. Similar struc-

Table 1. Discharge Rates of Rivers to the South Atlantic Bight

River	Rate of discharge, m ³ /s
Cape Fear	225
Pee Dee	393
Santee	60
Cooper	422
Savannah	342
Ogeechee	67
Altamaha	390
Satilla	63
St. Mary's	19
St. John's	89

This table was adopted directly from *Menzel* [1993, Table 2.1]. The data were taken from 7 to 74 years of stream gauge records at each river.

tures also are found for S_2 and N_2 tidal constituents. Their amplitudes, however, are one order of magnitude smaller than that of the M_2 tide. Computed amplitudes and phases of three semidiurnal tidal constituents are in good agreement with observational data available at Fernandina Beach (C_1), Florida, Savannah (C_2), Georgia, and Charleston (C_3), South Carolina. The differences of computed and observed amplitudes and phases at three stations are less than 2 cm and 1° , respectively, in the M_2 tide, 0.9 cm and 2° , respectively, in the S_2 tide, and 0.6 cm and 1° , respectively, in the N_2 tide (Table 2).

The model provides a reasonable simulation of the fortnightly and monthly variations of semidiurnal tides in the inner shelf of the SAB. A direct comparison between measured and computed elevations at stations C_1 , C_2 , and C_3 is shown in Figure 3 for a period of about 2 months. The measured data shown in Figure 3 are the envelopes of the maximum amplitude of the tidal elevation constructed from six tidal constituents (semidiurnal tides M_2 , S_2 , N_2 and diurnal tides O_1 , K_1 , P_1). A similar method is used to treat the computed tidal elevation constructed from three semidiurnal tidal constituents (M_2 , S_2 , and N_2). The envelopes of computed and measured tidal elevations coincide well each other over the fortnightly and monthly timescales. The small variation in the measured data was caused by diurnal tidal constituents that have not been included in the model simulation.

Synoptic distributions of computed M_2 tidal current vectors over a tidal cycle are shown with a time interval of 3 hours in Figure 4. The currents rotate clockwise, with the strongest cross-shelf transport in the inner shelf during flood and ebb tidal periods. The magnitude of tidal currents is strongest in the widest part of the shelf and decreases to the north and south where the shelf narrows. This spatial distribution of the semidiurnal tidal current is consistent with a simple theory of tidal waves proposed by *Clarke* [1991]. This theory

suggests that at latitudes where tidal frequencies are higher than the local inertial frequency, a large amplification of semidiurnal tides could occur over the wider continental shelf owing to a resonance mechanism.

Comparisons of measured and computed M_2 tidal currents at the 30-m and 40-m isobaths are presented in Figure 5. The magnitude and direction of computed velocity are in reasonable agreement with observations. The current meter data used in this comparison were digitized directly from *Wang et al.* [1984, Figures 5 and 6]. The small discrepancies found between computed and measured currents are possibly due to the temporal variation of internal and subtidal currents that were not completely removed in the 3- to 40-hour band-passed filtered current meter data.

Our model results of the M_2 tidal simulation are consistent with previous model efforts made by the vertically averaged model [*Wang et al.*, 1984], the diagnostic finite element model [*Werner et al.*, 1993], and the 3-D Princeton Ocean Model [*Kourafalou*, 1993]. Unlike previous works, our tidal model includes N_2 and S_2 tidal constituents. This allows us to simulate the fortnightly and monthly variations of semidiurnal tides. Although the effects of these tidal variations on tidally induced residual currents and tidal mixing over the inner shelf of the SAB are generally weak compared with the semidiurnal variation of the M_2 tide, they would play an important role in the temporal variation of water transport in estuaries. As a result of tidal simulation, we include them here and assume that they would be used for the development of the ecosystem model of Georgia estuaries in the future.

We also examined the influences of tidal forcing at open boundaries on tidal simulation in the inner shelf of the SAB by forcing the model with the M_2 tidal elevation adopted from *Schwiderski's* [1980] global $1^\circ \times 1^\circ$ tidal model. This model provides a reasonable result in terms of amplitude but not in phase, as do those predicted in the model forced by the global $0.5^\circ \times 0.5^\circ$

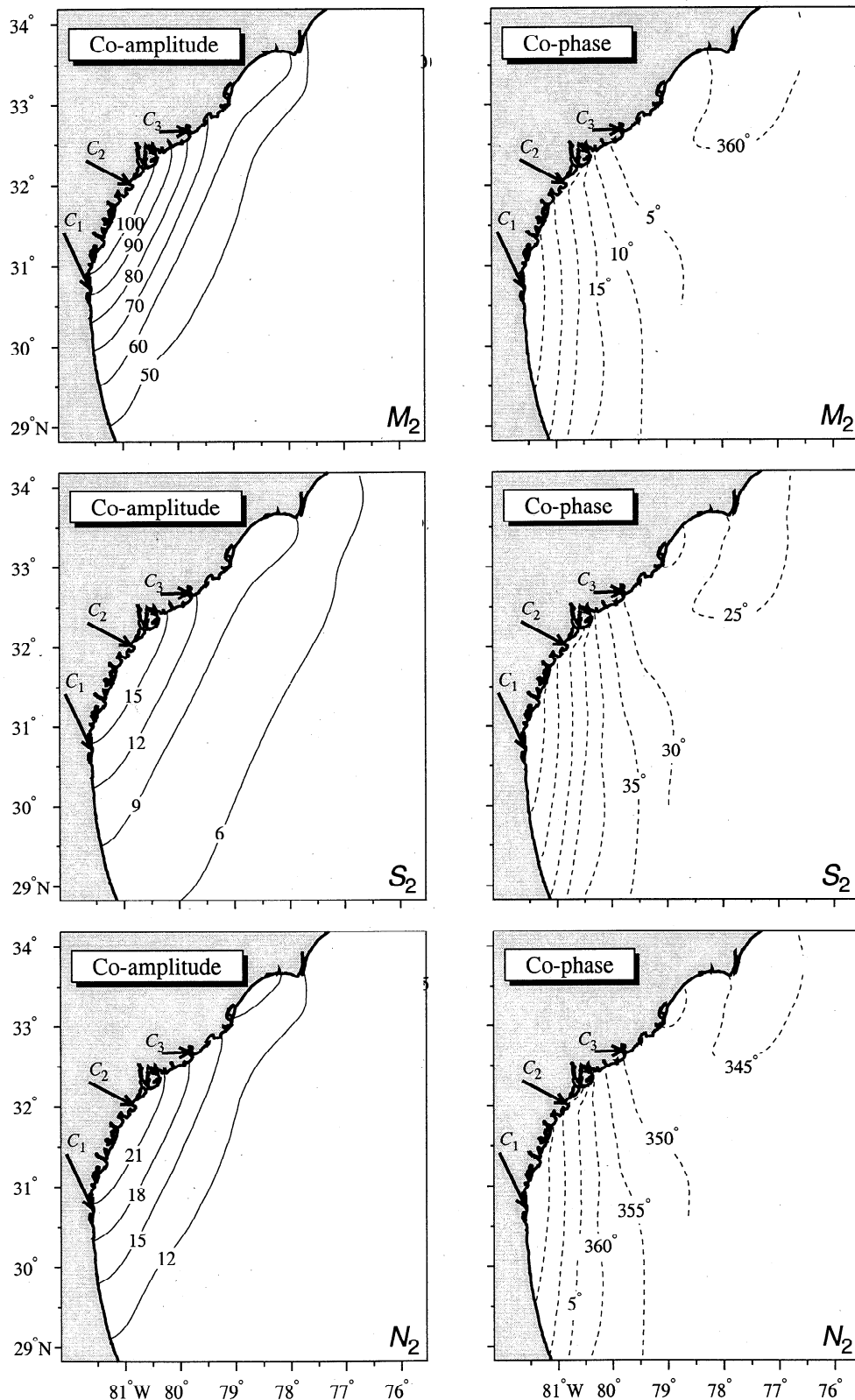


Figure 2. Computed coamplitudes (centimeters) and cophases (degrees Greenwich) of semidiurnal tidal constituents M_2 (12.42 hours), S_2 (12 hours), and N_2 (12.66 hours) in the inner shelf of the South Atlantic Bight (SAB).

Table 2. Comparison Between the Observed and Computed Amplitude and Phase of M_2 , S_2 , N_2 Tidal Constituents

Stations	Amplitude, cm						Phase, °Greenwich					
	M_2		S_2		N_2		M_2		S_2		N_2	
	ζ_o	ζ_c	ζ_o	ζ_c	ζ_o	ζ_c	θ_o	θ_c	θ_o	θ_c	θ_o	θ_c
Fernandina Beach	89.7	88.8	13.9	13.8	19.3	19.6	33.1	33.9	62.3	63.2	20.3	19.3
Savannah	101.7	103.3	16.7	16.5	22.0	22.6	17.5	17.5	45.9	47.4	4.9	4.9
Charleston	77.9	77.5	12.3	12.4	17.3	17.4	3.0	2.0	29.1	29.4	349.2	349.1

Fernandina Beach is located at $81^{\circ}27.6'W$, $30^{\circ}40.8'N$; Savannah is at $80^{\circ}54'W$, $32^{\circ}01.8'N$; and Charleston is at $79^{\circ}55.2'W$, $32^{\circ}46.8'N$. ζ_o : observed amplitude; ζ_c : computed amplitude; θ_o : observed phase; θ_c : computed phase.

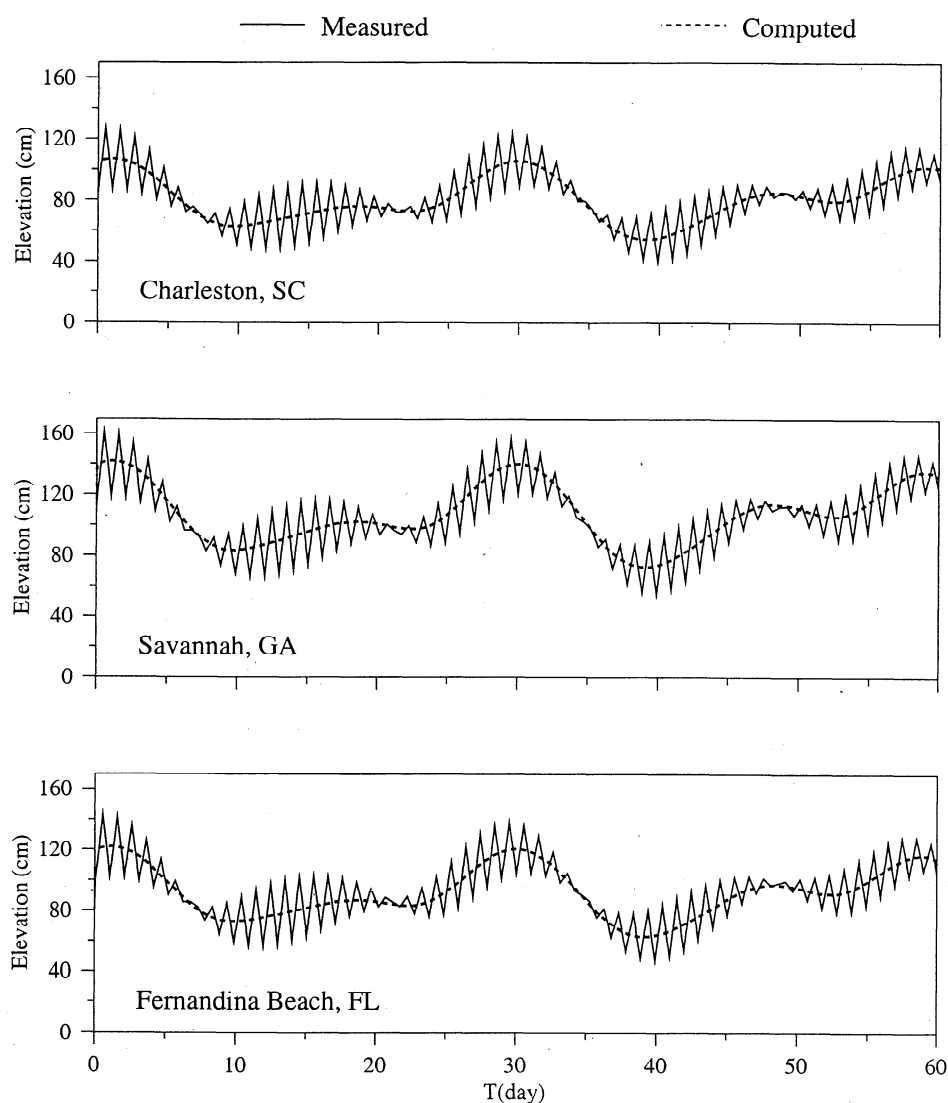


Figure 3. Comparison between computed and observed tidal elevations over a period of 2 months at tidal stations at Charleston, South Carolina, Savannah, Georgia, and Fernandina Beach, Florida. The observed data shown were the envelopes of the maximum amplitude of the tidal elevation constructed from six tidal constituents (semidiurnal tides M_2 , S_2 , N_2 and diurnal tides O_1 , K_1 , P_1). A similar method was used to treat the computed tidal elevation constructed from three semidiurnal tidal constituents (M_2 , S_2 , and N_2).

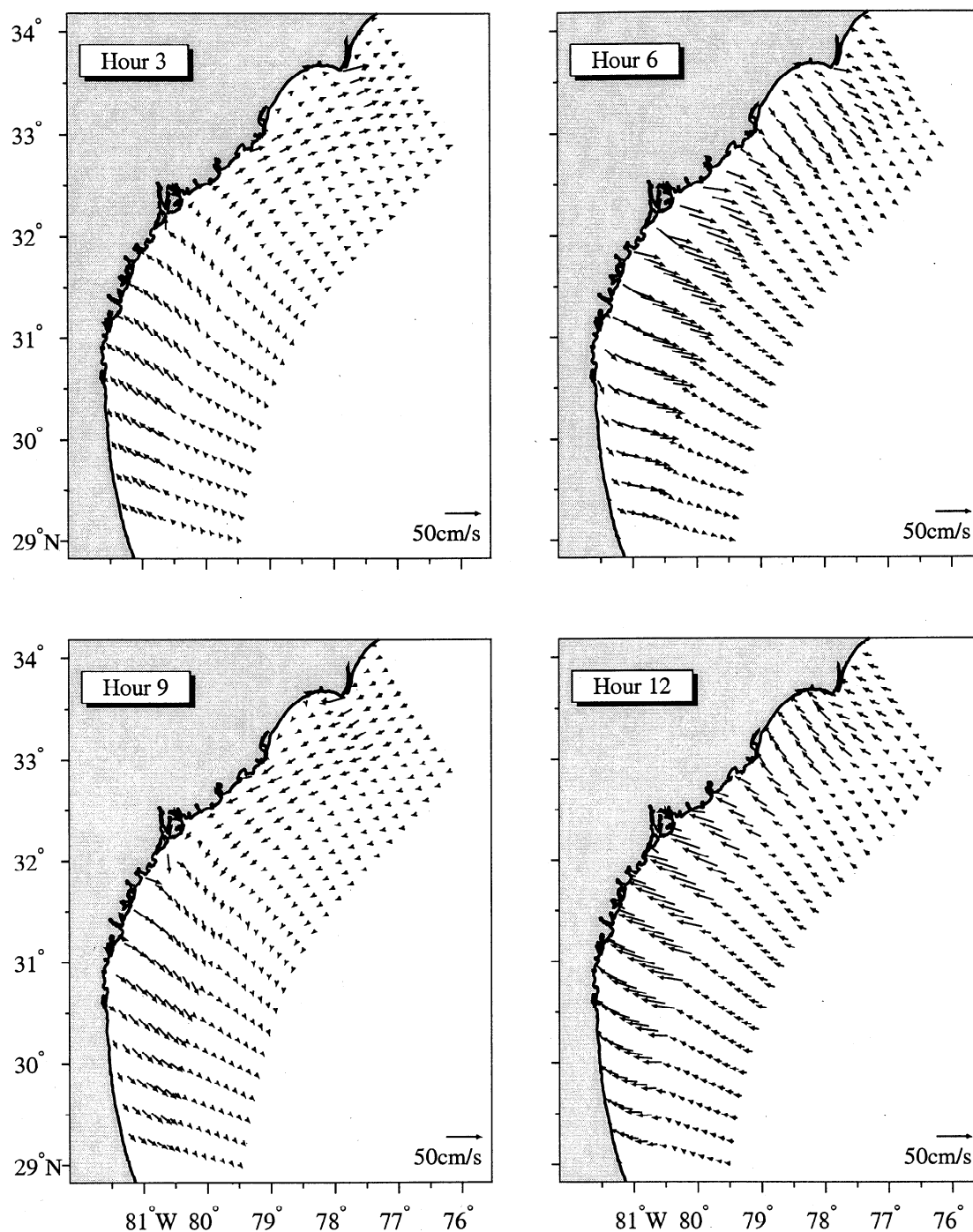


Figure 4. Synoptic distributions of the surface tidal current vectors of the M_2 tide at hours 3, 6, 9, and 12 over a tidal cycle. The vectors were plotted every five grid points.

tidal elevations. This suggests the importance of tidal forcing at open boundaries in tidal simulation over the inner shelf of the SAB.

To examine the tidal rectification processes over the inner shelf of the SAB, we have run the tidal simulation for a homogeneous case with only M_2 tidal forcing. Figure 6 shows the synoptic distribution of the near-surface residual current vectors in the center region of the inner shelf of the SAB. The model predicts a significant, topographically controlled, along-shelf residual

current near the Savannah River. This current moves northward along the coast, turns offshore at the inlet on the northern side of the river, and then flows southward along the 10-m isobath. Parts of the southward flow recirculate onshore to form an anticyclonic eddy near the mouth of the Savannah River. The maximum residual current is about 5 cm/s, occurring at the 10-m isobath offshore of the Savannah River. A weak anticyclonic eddy also is found near the Pee Dee River. It causes a weak coastal jet to flow northeastward at a speed of

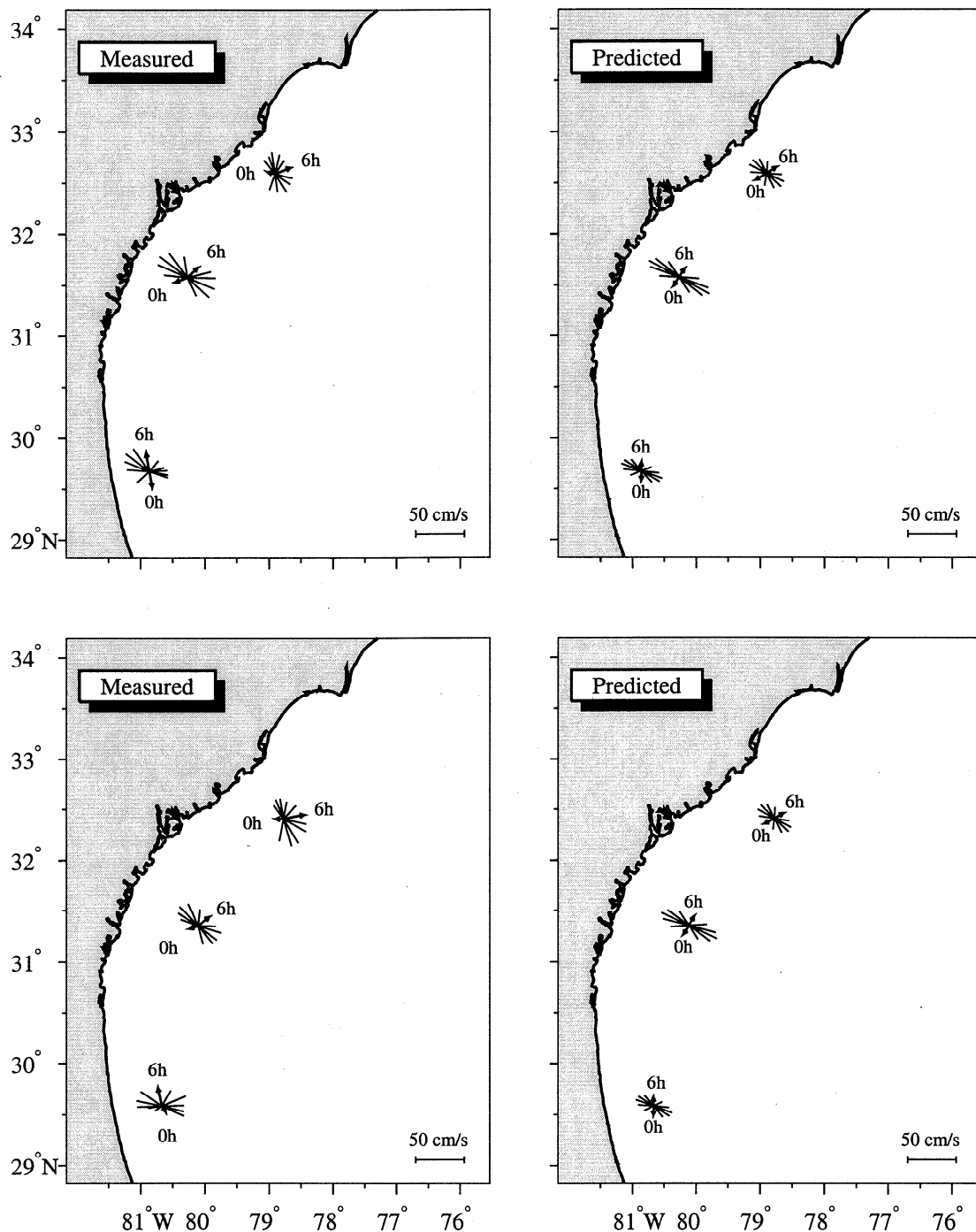


Figure 5. Comparison of measured and computed M_2 tidal currents at (top) 30-m and (bottom) 40-m isobaths. The measurement data used in this comparison were digitized directly from Wang *et al.* [1984, Figures 5 and 6].

about 2 cm/s and then turns clockwise at the crest of local isobaths.

Tongue-like shoals extend offshore in the region off the Cape Fear River. Interaction of tidal currents with this bottom topography generates an anticyclonic eddy between the 10- and 20-m isobaths, which swirls clockwise at a speed of about 2 to 3 cm/s (Figure 7).

It is difficult to use the previous current measurement data to verify the reality of model-predicted, tidally in-

duced residual currents near the coast, since rectified currents are relatively weak and all data available in autumn and winter were taken under a wind condition. A comparison between model-predicted and observed currents is made here in a qualitative sense. In the inner shelf region where the water depth is less than 10 m, strong mixing causes wind energy to transfer more quickly to the bottom, and the current responds instantaneously to local wind stress with a frictional adjust-

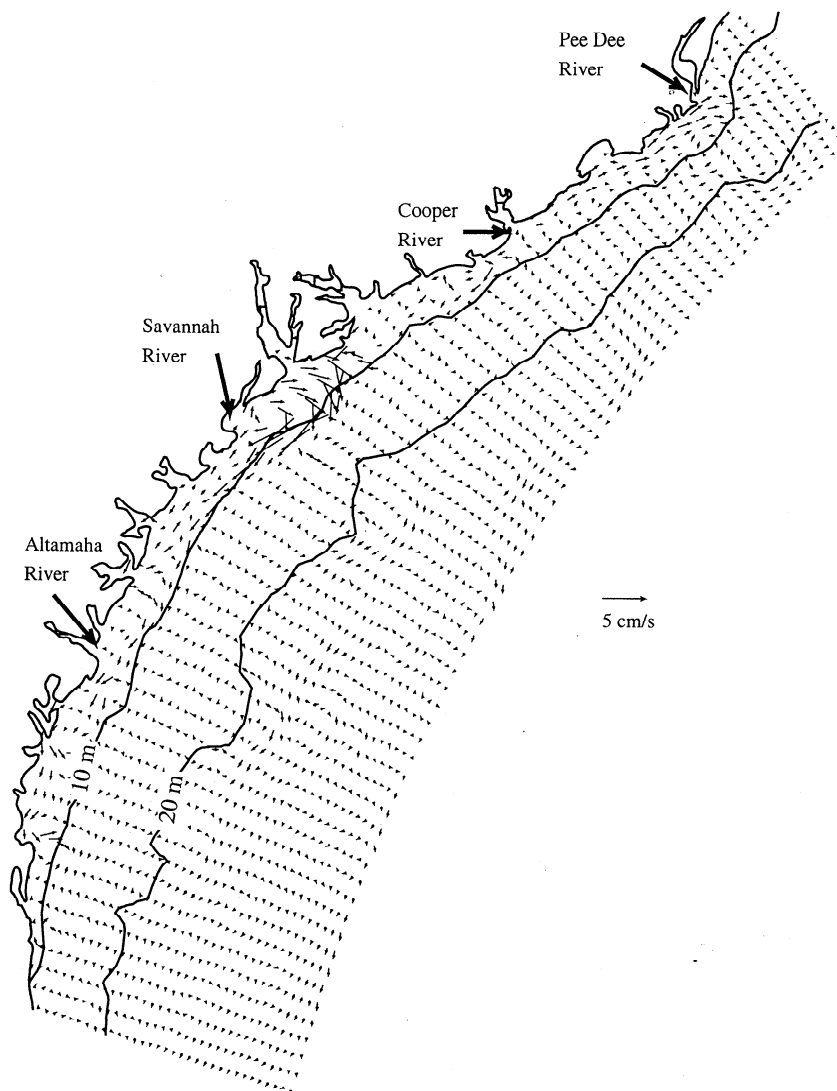


Figure 6. Distribution of the surface residual current averaged over a tidal cycle at the end of the 5th day. In this case, the model was forced only by the M_2 tidal elevation at open boundaries.

ment time of about 3.3 hours [Lee *et al.*, 1989]. In this case, the wind-induced, along-shelf current u_w can be estimated by a balance relation between wind and bottom stresses, which is given in the form of

$$u_w = \sqrt{\frac{|\tau_{sx}|}{\rho C_d}} \left(\frac{\tau_{sx}}{|\tau_{sx}|} \right) \quad (11)$$

During the Genesis of Atlantic Lows Experiment (GALE), current measurements were conducted at the 10-m isobath near the Pee Dee River from January to March 1986. When an along-shelf wind blew along the coast during January 22–23, a vertically uniform, northeastward along-shelf current was observed at mooring station 1 (see Figure 1). The magnitude of the current was about 10 cm/s under a wind stress of 0.3 dyn/cm² [Lee *et al.*, 1989]. Using the bottom drag coefficient value of $C_d = 5.57 \times 10^{-3}$ (estimated by Lee *et al.* [1989] based on the observed data), we can use (11) to estimate the wind-induced component of the along-shelf

current, which was equal to about 7 cm/s. Subtracting this wind-induced current component from the observed current yields a northeastward residual current of 3 cm/s. This residual current was probably driven by the cross-shelf gradient of surface elevation or tidal rectification. Our model predicts a northeastward, tidally rectified current of 1.4 cm/s at that location, which is in the same direction as the observed residual current. Similar results also were found near the Savannah River during Autumn 1987, where a southwestward along-shelf current of about -23 cm/s was observed at mooring station 2 (on the 8-m isobath) under a southward along-shelf wind stress of -2 dyn/cm² during late October [Blanton *et al.*, 1994]. The wind-induced current component estimated by (11) was about -19 cm/s. This suggests the existence of a southward current of about 4 cm/s that may not be driven by wind forcing. Our model predicted a southward tidally rectified current of about 5 cm/s at the same location near the Savannah River, which is the same in both magnitude and direction as the observed residual current.

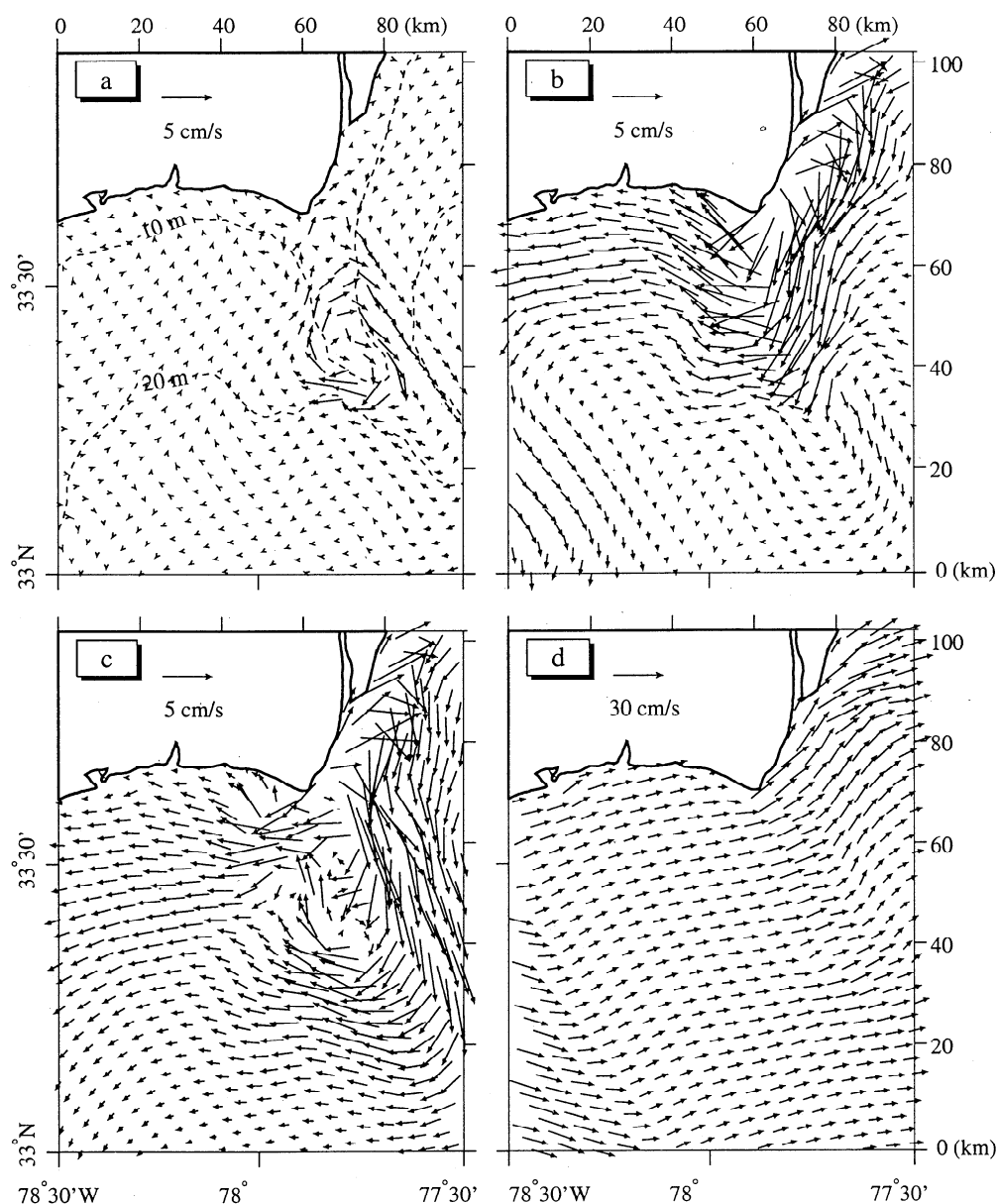


Figure 7. Distributions of the surface current near the Cape Fear River, averaged over a tidal cycle (a) at the end of the 5th day; (b) at the end of the 110th day for case with the only river discharges; (c) the M_2 tide plus river discharges; and (d) the M_2 tide, river discharges, and upwelling wind.

Dynamics of tidal rectification over a variable bottom topography have been well explored in many coastal regions [Huthnance, 1973; Zimmerman, 1978; Loder, 1980; Chen and Beardsley, 1995; Chen et al., 1995]. As a barotropic semidiurnal tidal wave propagates onto the shelf, rotation, mass continuity, and friction can cause the cross-shelf momentum flux of along-shelf tidal current to diverge on the side of the positive gradient bottom slope and to converge on the side of the negative gradient bottom slope. As a result, an along-shelf residual current forms over the shelf with a shallower region to the right [Loder, 1980]. The magnitude of this current depends on the magnitude of tidal currents and the slope of the bottom topography. This theory can be applied to the center region of the SAB between

Georgia and South Carolina, where the motion is dominated by tidal currents. Tidal currents are strongest near the coast in this region, which causes a stronger tidally averaged nonlinear advection near the Savannah River (Figure 8, left). On section 1, the along-shelf Coriolis force is mainly balanced by nonlinear advection, surface pressure gradient, and vertical diffusion. A relatively strong, southward along-shelf residual current found near the Savannah River is generated by nonlinear advection against friction, the same driving mechanism as Loder [1980] proposed. A clockwise residual circulation pattern there forms owing to the local bottom topography. A shallower 3-D sill exists near the mouth of the Savannah River, which drives a rectified flow clockwise with a shallower region to the right.

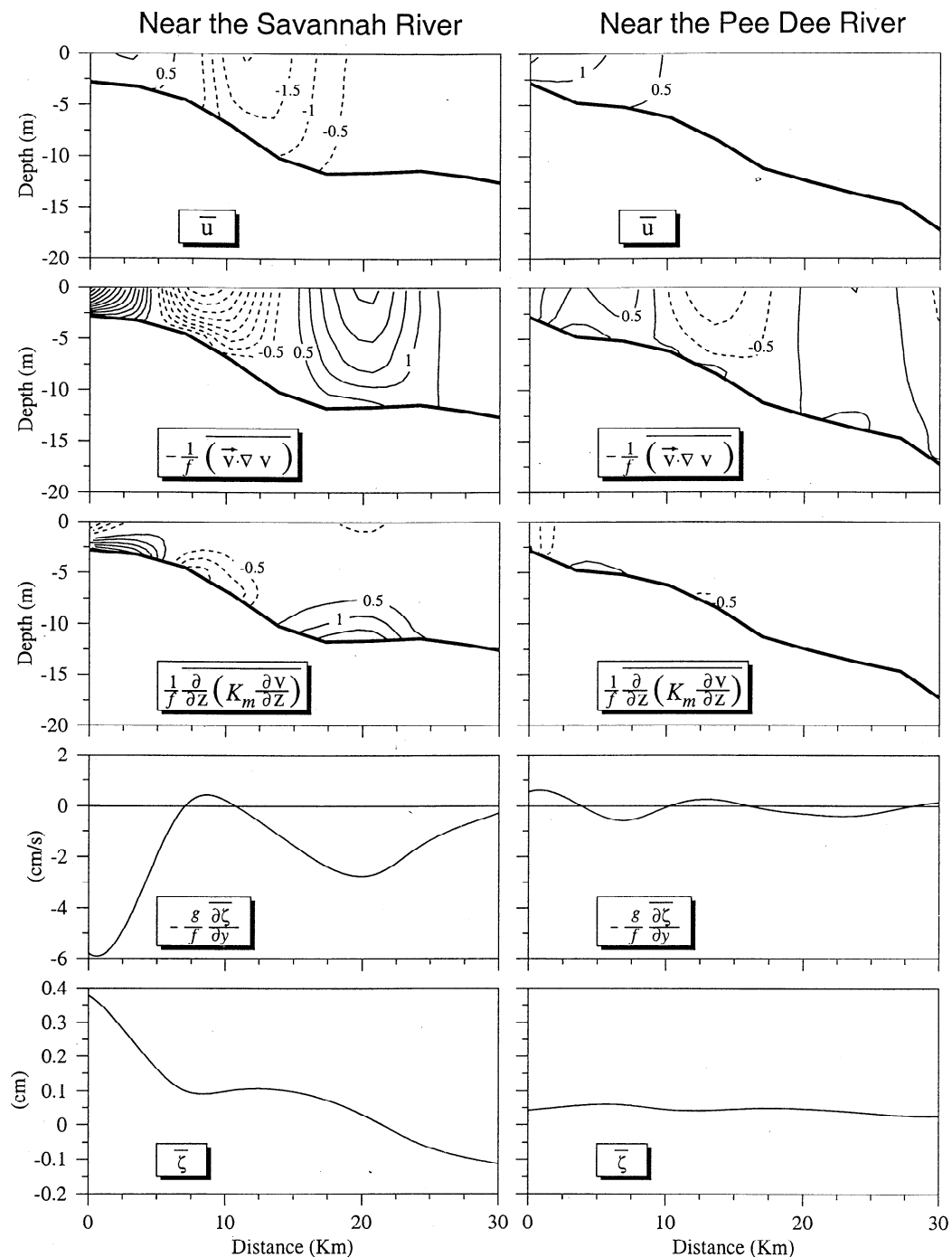


Figure 8. Distributions of the cross-shelf momentum averaged over a tidal cycle at the end of the 5th day on sections 1 (near the Savannah River) and 3 (near the Pee Dee River). In this case, the model was forced only by the M_2 tidal elevation at open boundaries. Unit is centimeters per second. See text for variable definitions.

Near the Pee Dee River, however, tidal current is weak and the vertical diffusion term is negligible in the cross-shelf momentum balance except in the narrow region close to the coast. A northward, along-shelf residual current found near the Pee Dee River is driven by the tidal advection and cross-shelf surface pressure gradient. The direction of this current is opposite to that

caused by tidal rectification shown by *Loder* [1980] and *Chen and Beardsley* [1995].

4. Formation of the Low-Salinity Front

To examine the effects of tidal mixing on the formation of the low-salinity front, we ran the model for two

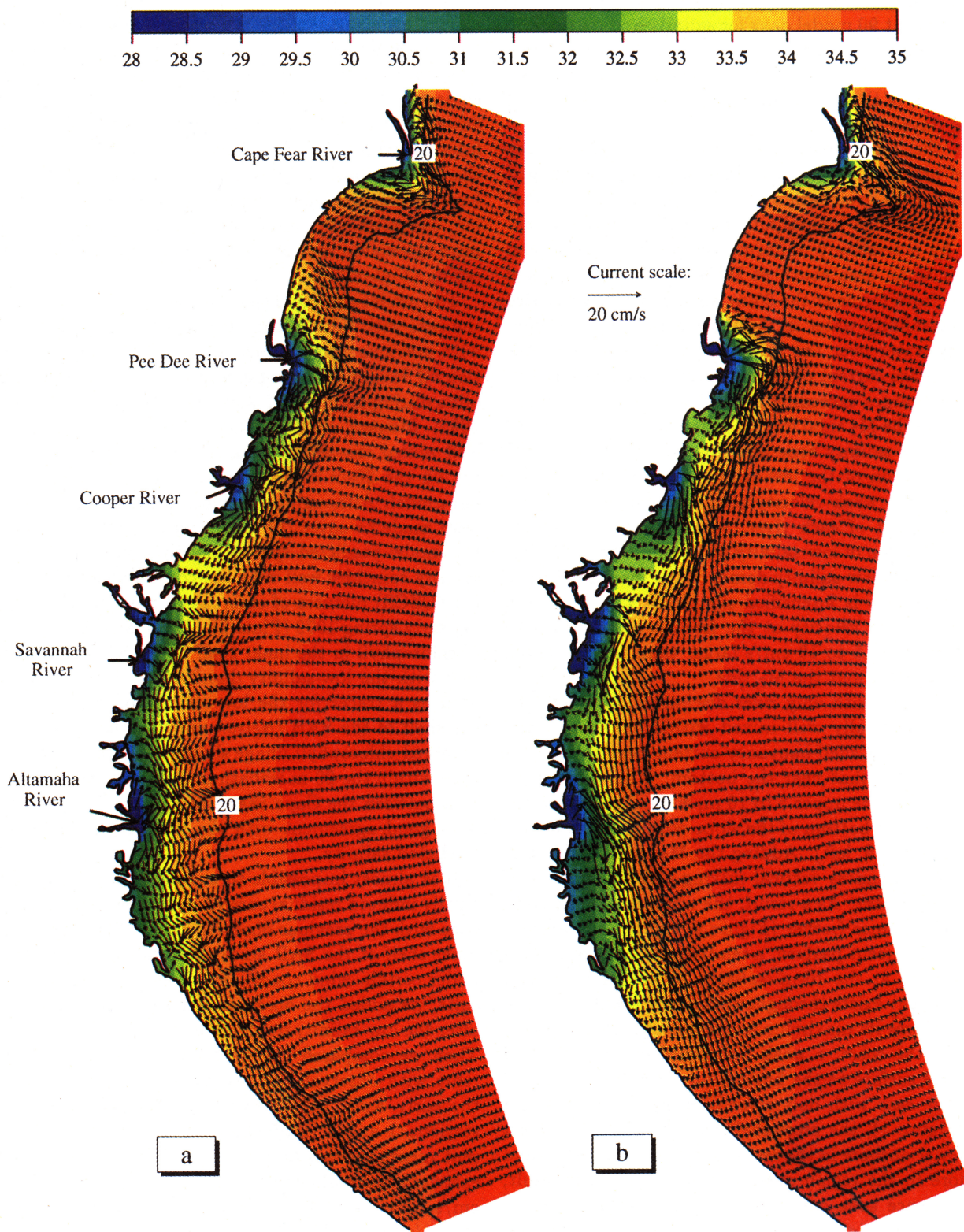


Plate 1. Distributions of the surface salinity and surface current vectors at the end of the 110th day for the cases (a) without and (b) with tidal forcing. In both the cases, the freshwater discharges were injected into the shelf from 10 rivers along the coast. The discharge rate from each river is given in Table 1.

cases: one was forced only by river discharges, and the other was forced by river discharges plus the M_2 tide. It took about 90 model days for salinity and currents to reach an equilibrium state in the first case, while it took almost 100 model days in the second case. To compare the results of these two cases, we ran the model for 110 model days.

For the case forced only by river discharges, the model predicts a reduced salinity zone that is trapped within the 20-m isobath over the inner shelf of the SAB (Plate 1a). The low-salinity water is advected southward by a relatively uniform, along-shelf velocity and arrives at an area south of 29°N on the 110th model day. A front forms at the 20-m isobath, which separates the low-salinity water in the inner shelf from the high-salinity water over the midshelf. A maximum southward surface current of about 10 cm/s occurs off rivers, with discharges exceeding $220 \text{ m}^3/\text{s}$.

Adding the M_2 tidal forcing causes an offshore shift of the reduced-salinity zone and slows or stalls the southward intrusion of the low-salinity plume (Plate 1 and Figure 9). A weak velocity area is found in the downstream region of each river, which blocks the southward advection of low-salinity water and forms multiple "tongue-like" low-salinity domes in the inner shelf. The frontal current is intensified significantly in this case. The maximum velocity at the front is about 15 to 20 cm/s, which is 5 to 10 cm/s larger than that observed in the case without the M_2 tide.

In the case with only river discharges, although the water is vertically well mixed within the inner shelf zone, the along-shelf buoyancy current near the surface is opposite to that near the bottom. The water flows southward in the upper 5 to 10 m, while it moves northward near the bottom of the slope within 10 km off the coast (Figure 9, left). Adding the M_2 tidal forcing significantly enhances vertical mixing and the cross-shelf salinity gradient. As a result, the water inside the 20-m isobath becomes fresher (Figure 9, right). Instead of a two-layer current pattern, a relatively strong southward jet appears at the 10-m isobath, 10 km off the coast. This current exists throughout the whole water column and achieves its maximum at the surface. In the region close to the coast, a weak northward current is found from the surface to the bottom on the 5-m isobath.

In the case with only river discharges, on section 1 (see Figure 1), the cross-shelf current tends to advect the water onshore in the region shallower than 12 m and offshore in the region deeper than 15 m. As a result, a divergence zone occurs on the 13-m isobath, 20 km off the coast (Figure 9, left). When the tide is included, the maximum divergence zone shifts to the center of the southward residual current jet (Figure 9, right). On the offshore side of the jet, the water is advected offshore near the surface by a strong current and reversed near the bottom by a weak current. On the onshore side of the jet, the cross-shelf residual flow is dominated by the onshore currents. Their magnitudes are, however,

much weaker than those found in the case forced only by river discharges.

The tide also intensifies the topographically controlled, residual eddy-like circulation off the Cape Fear River. The low-salinity water originating from the river travels in two branches in the case with tidal forcing: one flows southward along the coast, and the other circulates following a clockwise loop that has its center at the 20-m isobath (Figure 7c). The clockwise residual loop current disappears in the case without tidal forcing, suggesting the importance of the physical mechanism associated with stratified tidal rectification in that region (Figure 7b).

The model-predicted spatial distribution of salinity in the case with tidal forcing is qualitatively in agreement with previous observations taken on the cross-shelf section off Savannah, Georgia, and Charleston, South Carolina during spring 1985 and 1993 [Nelson and Guarda, 1995; Blanton, 1996]. Those data show a vertically well mixed low-salinity front with an offshore boundary on the 20-m isobath, similar to that predicted by our model. Although no previous hydrographic surveys were made to cover the entire inner shelf of the SAB, the model-predicted spatial distribution of the surface salinity coincides well with the coastal zone color scanner (CZCS) pictures of chlorophyll concentrations during early spring 1979 (Plate 2). The spatial distribution of chlorophyll in the inner shelf of the SAB is coherent well with the low-salinity distribution, which could be used as a tracer for the low-salinity plume in the inner shelf of the SAB [Nelson and Guarda, 1995]. The observed chlorophyll image shows a high concentration zone within the 20-m isobath and a weak concentration area between the Cape Fear and Pee Dee Rivers, which is in good agreement with the distribution of model-predicted salinity.

It is almost impossible to compare model-predicted, tidally induced residual current in the low-salinity frontal zone with the previous observed residual currents in the inner shelf of the SAB since those data were all taken under certain wind conditions. To verify our model results, we added the observed wind stress to the model after the low-salinity front reached an equilibrium state. Then we compared the model-predicted residual currents with observations. An example is given here for the early April 1985 measurements taken at mooring stations 3 and 4 [Paffenhöfer *et al.*, 1994]. The current measurements at a depth of 5 m showed a north-eastward residual current of 20 cm/s (along shelf, 18.9 cm/s, and cross shelf, 6 cm/s) at mooring station 3 and 22 cm/s (along shelf, 21 cm/s, and cross shelf, 7 cm/s) at mooring station 4 under a northeastward wind stress of 1.45 dyn/cm^2 . The model-predicted residual currents under the same wind stress are about 17.12 cm/s (along shelf, 17 cm/s and cross shelf, 2 cm/s) and 22.14 cm/s (along shelf, 21 cm/s and cross shelf, 7 cm/s) at the same depth at mooring stations 3 and 4, which are in good agreement with observed currents.

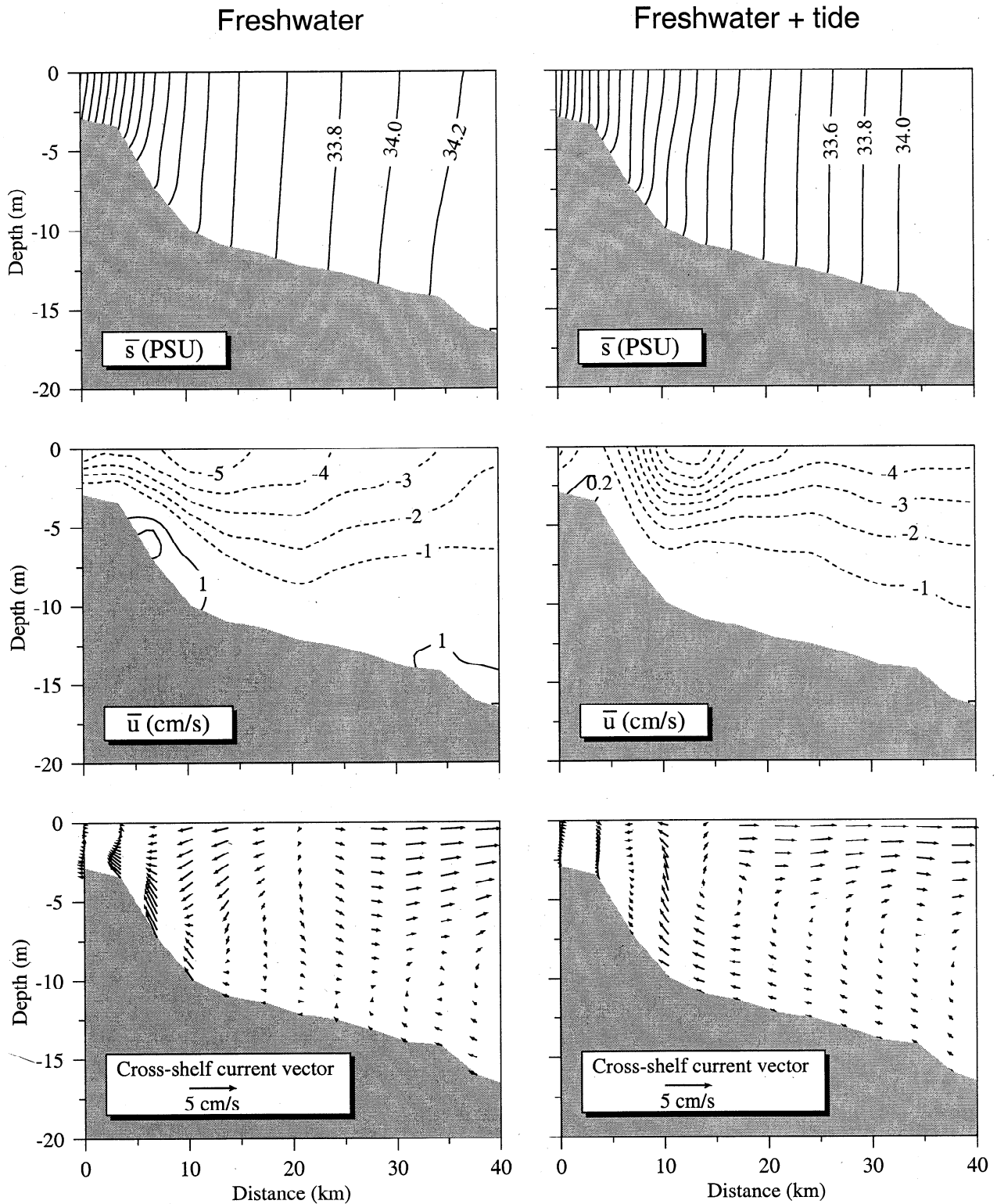


Figure 9. Distributions of (top) salinity, (middle) along-shelf current, and (bottom) cross-shelf current vectors averaged over a tidal cycle at the end of the 110th day on section 1, showing (left) freshwater discharges only and (right) freshwater discharges plus the M_2 tide.

Physical processes that control the southward intensified current at the front and northward reversed current near the coast are examined based on the momentum balances. After the low-salinity front reaches an equilibrium state, the along-shelf residual velocity is determined by

$$\bar{u} = -\frac{g}{f} \frac{\partial \zeta}{\partial y} - \frac{1}{f} \frac{\partial P}{\partial y} + \frac{1}{f} \frac{\partial}{\partial z} \left(K_m \frac{\partial v}{\partial z} \right) - \frac{1}{f} \mathbf{v} \cdot \nabla v \quad (12)$$

where the overbar indicates the mean value averaged over a tidal cycle, and P is the baroclinic pressure. In the case with only river discharges, the residual nonlin-

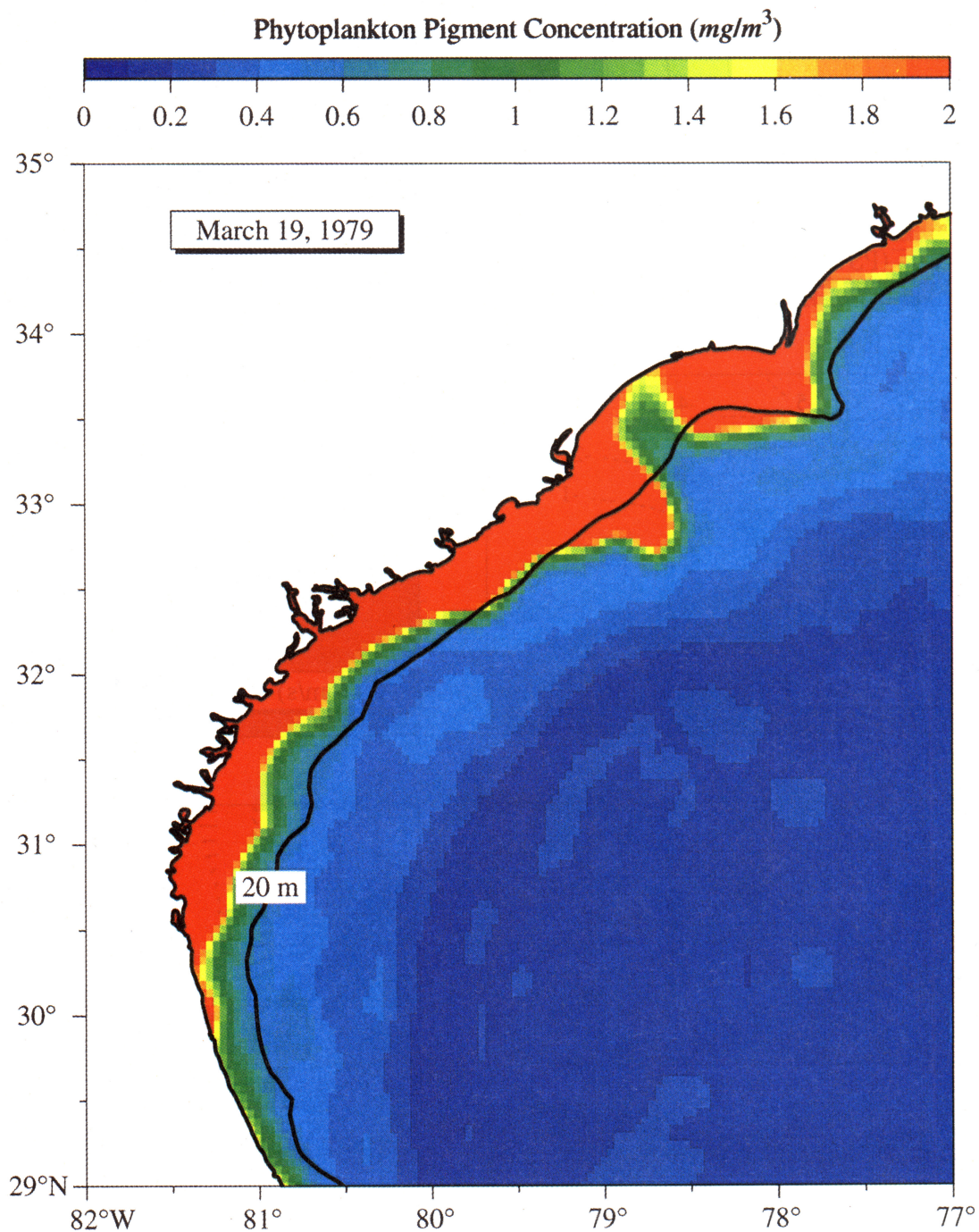


Plate 2. Image of the coastal zone color scanner (CZCS) chlorophyll concentrations on March 19, 1979. The solid line indicates the 20-m isobath.

ear advection terms are too small to be taken into account in the cross-shelf momentum balance (Figure 10, left). A negative surface pressure gradient force found near the coast tends to produce a barotropic, southward, along-shelf residual current. However, a positive baroclinic pressure gradient force, which is largest near the bottom and decreases upward, tends to cause a northward along-shelf current component. As a result, in the upper layer, the along-shelf residual current is southward and driven quasi-geostrophically by the surface pressure gradient and Coriolis forces. This southward quasi-geostrophic current is significantly reduced with depth and reverses near the bottom. The northward along-shelf residual current near the bottom is driven mainly by a positive baroclinic pressure gradient force balanced by the surface pressure gradient force, vertical diffusion, and Coriolis force.

When the M_2 tide is included, the magnitudes of nonlinear advection and vertical diffusion terms significantly increase, especially near the coast where the water depth is less than 10 m (Figure 10, right). Although the signs of surface and baroclinic pressure gradients remain unchanged, the cross-shelf distributions of these two terms are modified. The maximum magnitude of the surface pressure gradient force, which occurs about 3 km away from the coast in the case without tidal forcing, shifts to the coast after the tide is added. A peak of the surface pressure gradient force also is observed near the location of the maximum along-shelf residual current at the front in the case with tidal forcing.

Three physical processes have been recognized as playing critical roles in the intensification of residual current over a stratified shelf: (1) formation of the tidal mixing front, (2) stratified tidal rectification, and (3) modification of internal friction [Chen and Beardsley, 1995]. Over the inner shelf of the SAB, the cross-shelf salinity gradient is significantly enhanced due to tidal mixing and offshore extension of the low-salinity zone. The maximum difference in the baroclinic pressure gradient force terms (divided by f) between the cases with and without tidal forcing is about 0.7 cm/s near the surface, which is one order of magnitude smaller than nonlinear advection and vertical diffusion terms. This suggests that the modified density gradient is not primarily responsible for the intensification of the southward current at the front.

Stratified tidal rectification is a process associated with the nonlinear interaction between barotropic and internal tides as well as internal tides themselves over a variable bottom topography. In the inner shelf region, after the tide is added, the magnitude of the nonlinear advection term significantly increases to 10.5 cm/s near the coast and to -4.2 cm/s about 7 km off the coast. As a result of stratified tidal rectification, the southward residual current is significantly reduced near the coast and is remarkably enhanced on the slope between 5 and 10 km off the coast.

The modification of internal friction and the vertical shear of the residual current causes a significant increase of the vertical diffusion term near the surface and bottom. The maximum value of this friction term is about -19.0 cm/s near the bottom 5 km off the coast, 7.5 cm/s near the surface 7 km off the coast, and -3.3 cm/s near the surface at the center of the southward current jet. This is the same order of magnitude as the nonlinear advection term. A negative vertical diffusion, which occurs near the surface on the 10-m isobath, directly enhances the southward residual current jet.

Adding the tidal forcing causes a relatively large, negative cross-shelf surface pressure gradient force at a location 12 km away from the coast. This location coincides well with a center of the southward jet current, suggesting that the cross-shelf surface pressure gradient force plays an important role in intensifying the southward current at the front.

These analyses suggest that the intensification of the southward residual flow at the front and the weakly northward residual flow near the coast are mainly caused by stratified tidal rectification, modified internal friction, and the cross-shelf barotropic surface pressure gradient force.

The secondary cross-shelf residual circulation is controlled by tidally averaged along-shelf surface and baroclinic pressure gradients, vertical diffusion, and nonlinear advective terms given as

$$\bar{v} = \frac{g}{f} \frac{\partial \bar{\zeta}}{\partial x} + \frac{1}{f} \frac{\partial \bar{P}}{\partial x} - \frac{1}{f} \frac{\partial}{\partial z} \left(K_m \frac{\partial u}{\partial z} \right) + \frac{1}{f} \mathbf{v} \cdot \nabla u \quad (13)$$

In the case without tidal forcing, the nonlinear advective terms are too weak to be taken into account in the momentum balance (Figure 11, left). The along-shelf surface pressure gradient term is largest near the coast and decreases offshore with a sign change on about the 12-m isobath. The cross-shelf distribution of this surface pressure gradient term tends to cause a divergent flow in the cross-shelf direction, flowing onshore in the region with a positive sign of $\partial \zeta / \partial x$ and offshore in the region with a negative sign of $\partial \zeta / \partial x$. The baroclinic pressure gradient term is negative throughout the water column, with a maximum at the bottom near the coast. This term tends to produce an offshore component of the residual current. The vertical diffusion term is characterized by a two-layer structure, with a negative maximum value at the surface and a positive maximum value at the bottom. This frictional term tends to enhance the offshore current near the surface and the onshore current near the bottom. The weak secondary cross-shelf divergent residual flow pattern predicted by our model is driven by a resultant force of these three dynamical terms balanced with the Coriolis force.

When the tidal forcing is included, tidal mixing and stratified tidal rectification significantly increase the magnitude of vertical diffusion, nonlinear advective terms,

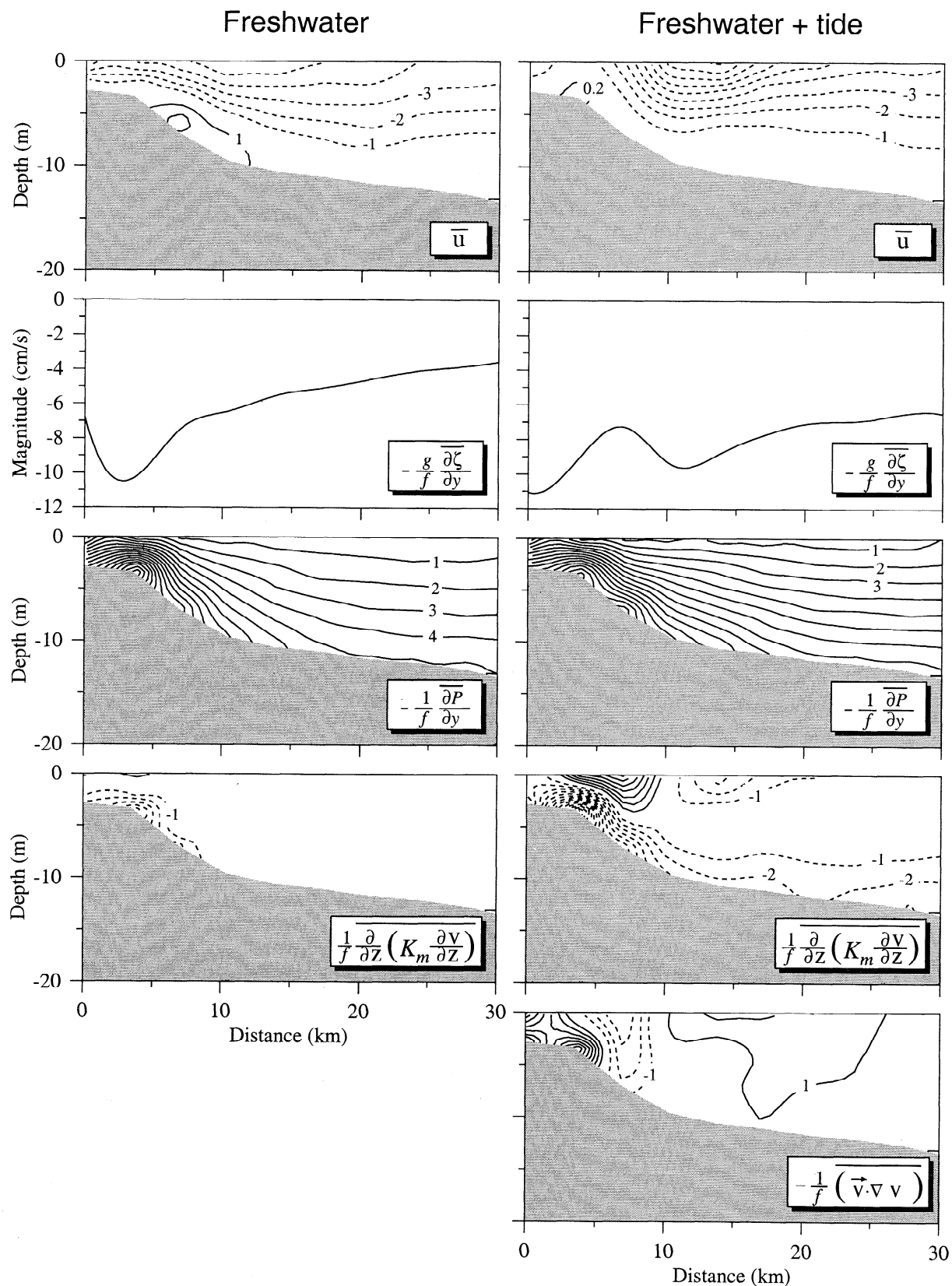


Figure 10. Distributions of the cross-shelf momentum averaged over a tidal cycle at the end of the 117th day on section 1, showing (left) freshwater discharges only and (right) freshwater discharges plus the M_2 tide. Unit is centimeters per second.

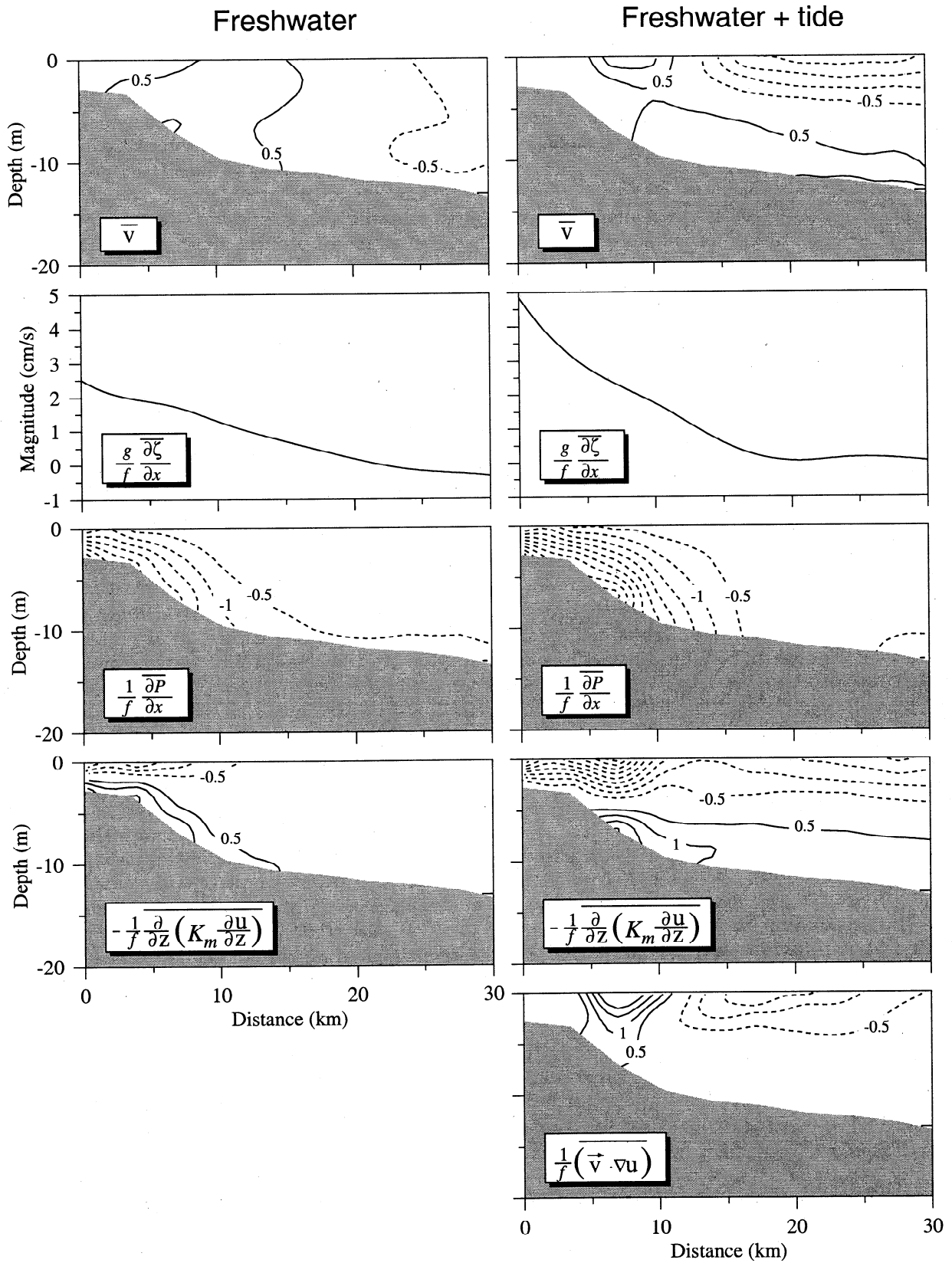


Figure 11. Distributions of the along-shelf momentum averaged over a tidal cycle at the end of the 117th day on section 1, showing (left) freshwater discharges only and (right) freshwater discharges plus the M_2 tide. Unit is centimeters per second.

and surface and baroclinic pressure gradients (Figure 11, right). Although the secondary cross-shelf residual circulation is still characterized by a divergent flow near the surface and convergent flow near the bottom, the strength of the near-surface current is much stronger. The center of the near-surface divergent zone shifts to the location of the maximum southward residual current jet.

The pattern of the secondary cross-frontal residual circulation, which diverges near the surface and converges near the bottom, was first revealed in previous modeling studies of low-salinity fronts by *Chapman and Lentz* [1994] and also in modeling simulation of tidal mixing fronts by *Chen et al.* [1995]. The formation of this circulation pattern is related to turbulent mixing in the bottom boundary layer. A detailed discussion of this issue is given by *Chapman and Lentz* [1994] and *Chen and Beardsley* [1998].

Chapman and Lentz [1994] used a time-dependent, 3-D numerical model to examine the dynamics of a surface-bottom density front on an idealized, uniformly sloping continental shelf. They found that the offshore buoyancy flux in the bottom boundary layer tends to move the front offshore until it reaches a depth where the vertical shear within the front produces a reversed cross-shelf velocity at the shoreward edge of the front. A steady state can be established in the downstream direction, where the vertical diffusion of density is balanced by the cross-shelf advection of density. Our model results obtained from the case forced only by multiple river discharges are very similar to those found by *Chapman and Lentz* [1994]. We did find a quasi-steady front farther downstream after 110 model days that is balanced by the vertical diffusion and cross-shelf advection of salinity.

Unlike the single river discharge problem studied by *Chapman and Lentz* [1994], our model was forced by multiple river outputs with different discharge rates under the tidal environment. Since these rivers are close to each other and tidal mixing is spatially dependent, the dynamics of the low-salinity front in the central part of the inner SAB shelf are controlled by a complex nonlinear process. For the case with only constant river discharges, a quasi-steady front develops at about the 20-m isobath with a salt balance between horizontal advections, horizontal diffusion, and vertical salt diffusion (Figure 12, left). Vertical advection of salinity becomes significant after tidal forcing is added. At that point, the salinity balance is between the horizontal advections, vertical advections, vertical diffusion, and horizontal diffusion of salinity (Figure 12, right).

5. Formation of Isolated Low-Salinity Lenses

To study the physical mechanism of wind-induced, cross-frontal water exchange in the inner shelf of the

SAB, we forced the model by a constant upwelling-favorable wind. After the low-salinity front was generated, a southwesterly wind of 5 m/s was added. The model shows that the near-surface current was reversed to the northward direction a few hours after the wind started (Plate 3a). The interaction between buoyancy- and wind-induced currents under the tidal environment caused a relatively strong offshore transport of the low-salinity water off the mouth of each river. A geographically controlled wave-like shape of salinity contours developed at the edge of the front in the middle shelf 8 model days after the wind blew. Contour crests occurred near the St. Johns, Savannah, Altamaha, and Cooper Rivers where the river discharges were larger (Plate 3b). Isolated lenses then formed at crests after 12 model days (Plate 3c). The detachment of isolated low-salinity lenses also were evident in Figure 13, which shows the time sequence of the offshore movement of the low-salinity water on section 2 across the crest near the Altamaha River. The low-salinity water in the upper 5 m at the crest was advected offshore at a speed of about 5 km/d during the first 10 model days. An isolated lens, marked with a contour of 34.6, occurred near the surface at the outer edge of the front on the 12th model day. This lens was completely separated from the front by the 14th model day when the contour of 34.8 outcropped the surface.

What are the physical processes responsible for the formation of isolated low-salinity lenses over the inner shelf of the SAB? Is it wind-induced upwelling, along-shelf advection, or vertical salt diffusion? A simple conceptual Ekman transport theory suggests that the response of water currents to wind varies from the inner shelf to middle and outer shelves. In the inner shelf, where the water depth is shallow and friction is large, the current responds almost instantaneously to local wind stress. As a result, water moves in the same direction as the wind. In the middle and outer shelves, however, the water is relatively deep and friction is weak, so the wind-driven current follows the Ekman theory. An upwelling-favorable wind tends to produce an offshore Ekman transport near the surface and an onshore Ekman transport near the bottom. Upwelling must occur at a place near the outer edge of the inner shelf in order to compensate for the offshore loss of the water near the surface. A low-salinity isolated lens can be formed as the outcropping tendency of isopycnals driven by upwelling. [*Csanady*, 1974; *Oey*, 1986; *Blanton et al.*, 1989]. The simple conceptual model, however, is built for a 2-D case in which the wind is the only forcing. This model fails to explain why the isolated lenses occur only in some particular regions over the inner shelf of the SAB.

Figure 14 shows the time sequence of the cross-shelf distribution of near-surface residual current vectors on section 2. The currents completely reversed to the north 1 day after the wind blew. The wind-induced northward

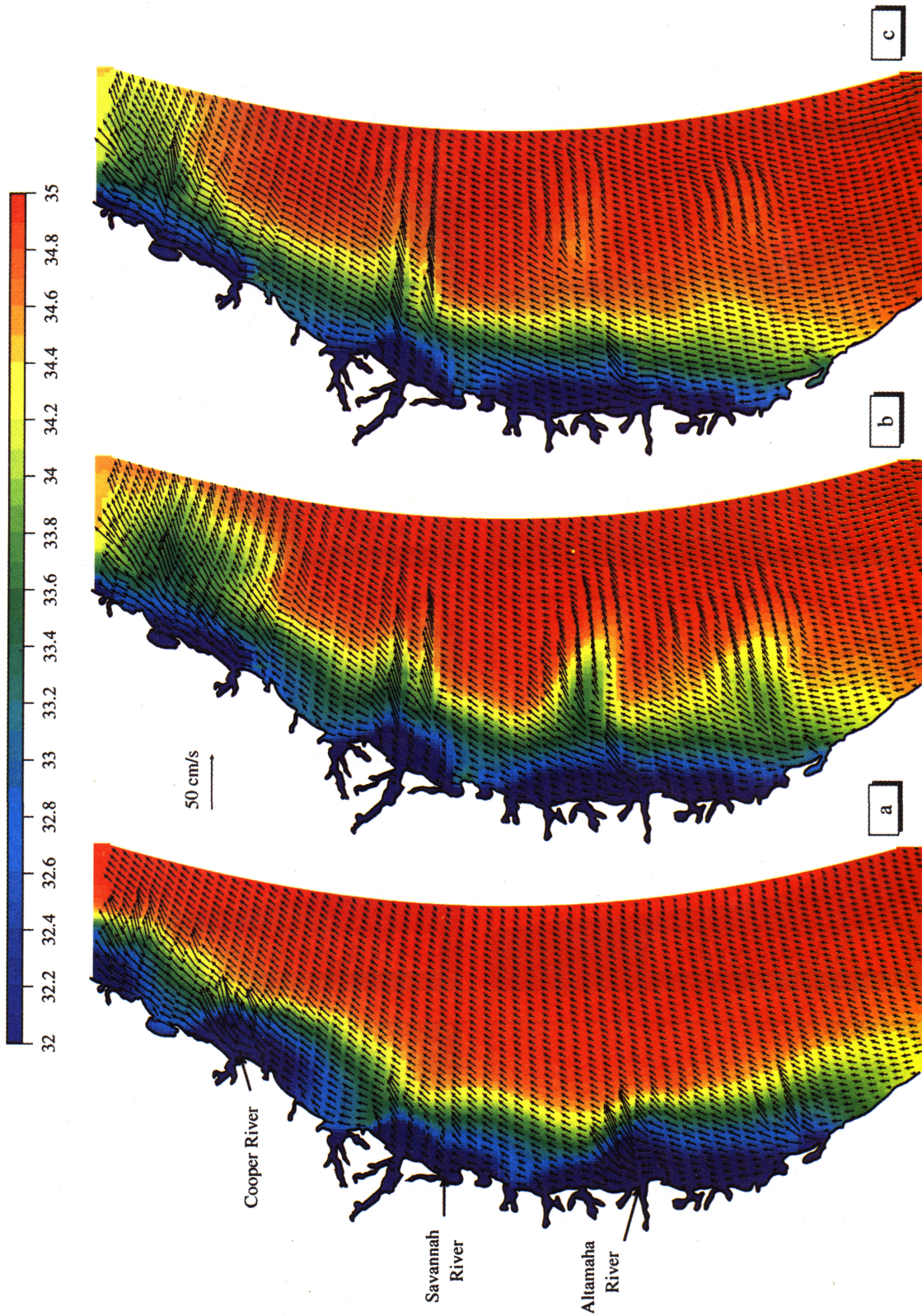


Plate 3. Distributions of surface residual salinity and current vectors averaged over a tidal cycle on days (a) 1, (b) 8, and (c) 12 after the wind blew. In this case, the model was initially forced by multiple river discharges plus M_2 tide as shown in Plate 1 (b). A constant upwelling-favorable wind of 5 m/s was added at the beginning of the 11th day.

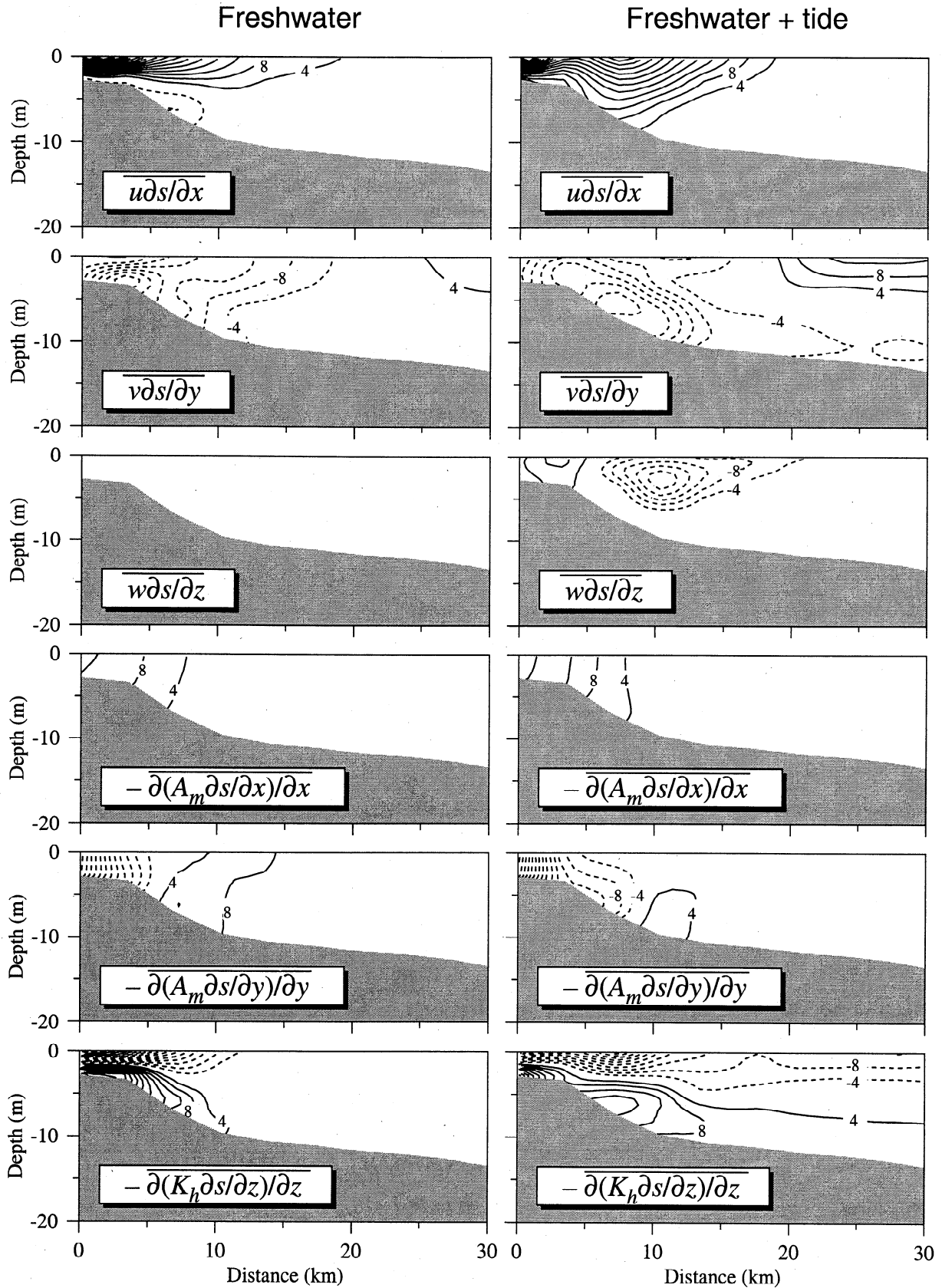


Figure 12. Distributions of salt advection and diffusion terms in the cases with (left) only river discharges and (right) river discharges plus the M_2 tide at the end of the 110th day on section 1. The terms in Figure 12 right, were the residual values averaged over a tidal cycle. Unit is 10^{-7} psu/s.

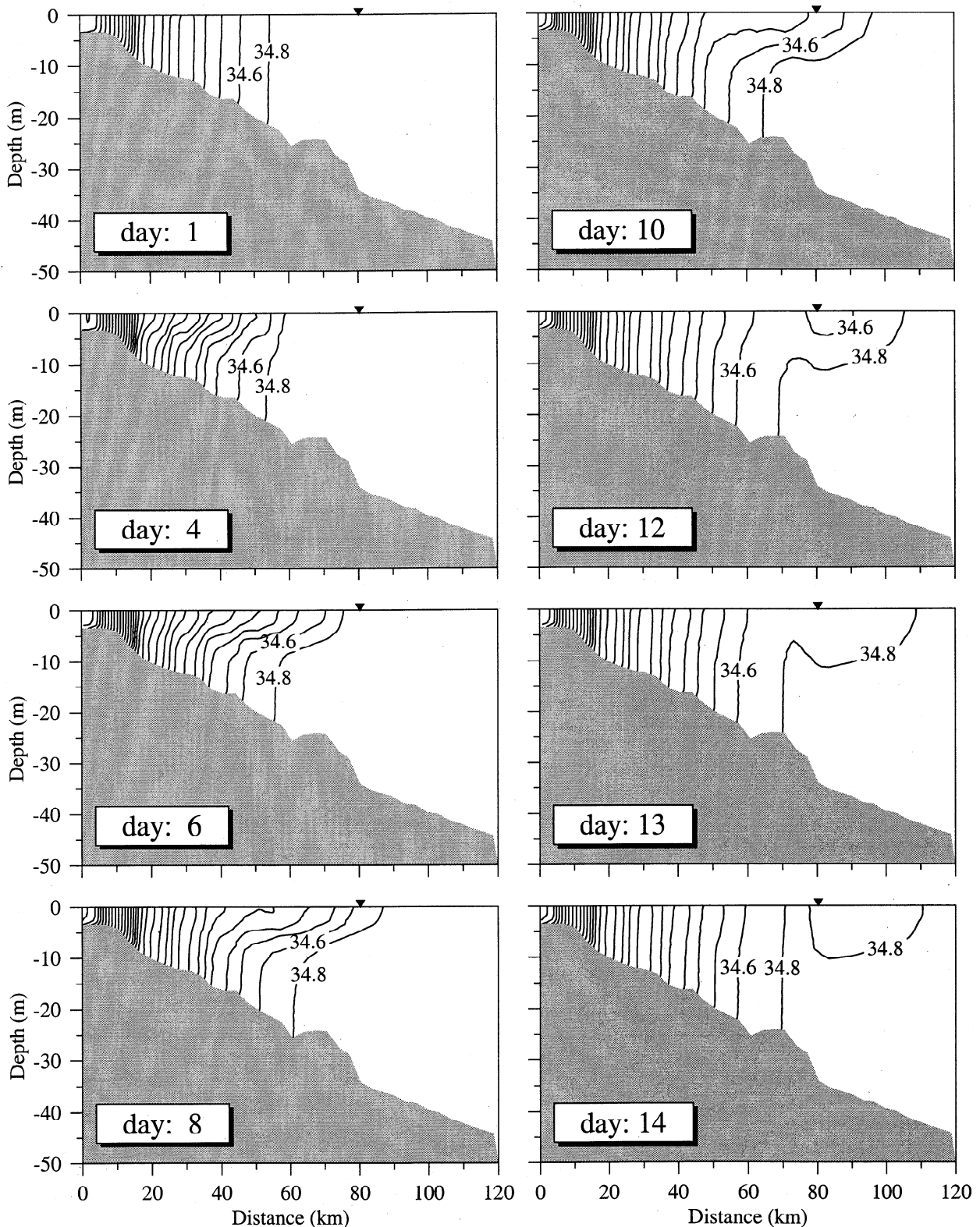


Figure 13. Cross-shelf distributions of the salinity averaged over a tidal cycle on section 2 from day 1 to day 14 after the wind blew. The solid triangles indicate the detachment point of the low-salinity water at the edge of the front.

currents within the frontal zone, between 20 and 60 km, were significantly enhanced on the 4th model day. The current at the outer edge of the front tended to turn clockwise, pushing the low-salinity water offshore. The offshore transport process continued until a northward

current was reestablished between 60 and 80 km on the 12th model day.

Physical processes that cause the increase of the offshore water transport at the crest of the outer frontal edge can be estimated by a transport model derived

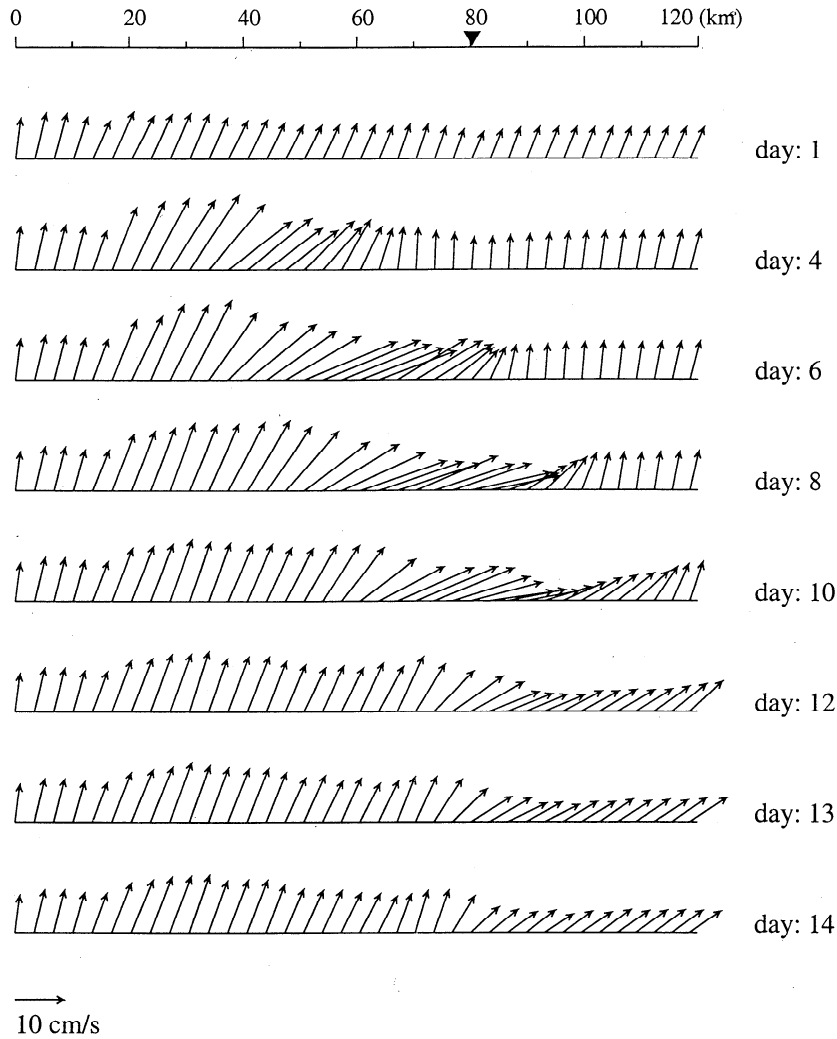


Figure 14. Time sequences of the cross-shelf distribution of the surface residual current vectors from day 1 to day 14 after the wind blew. The solid triangle indicates the detachment point of the low-salinity water at the edge of front.

from the momentum and continuity equations. This model is given as

$$\begin{aligned} & \left(\int_{-d}^{\zeta} v dz \right)_y = \\ & \left(\int_{-d}^{\zeta} v dz \right)_{y_0} + \frac{g}{f} (\zeta + d) \left[\left(\frac{\partial \zeta}{\partial x} \right)_y - \left(\frac{\partial \zeta}{\partial x} \right)_{y_0} \right] + \\ & \frac{1}{f} \int_{-d}^{\zeta} \left[\left(\frac{\partial P}{\partial x} \right)_y - \left(\frac{\partial P}{\partial x} \right)_{y_0} \right] dz + \frac{1}{f} [(\tau_{dx})_y - (\tau_{dx})_{y_0}] \quad (14) \end{aligned}$$

where y is the detachment location of the low-salinity water, y_0 is the station at the coast, d is the depth of the surface wind-induced mixed layer (it is assumed to be independent of y), and τ_{dx} is the stress at the base of the surface mixed layer. The offshore water transport of the mixed layer at the detachment location of low-salinity water significantly increased 4 days after the wind blew and reached a maximum at the 10th model day (Fig-

ure 15a). The increase of the offshore water transport was caused dominantly by the decrease of the second term on the right-hand side of (14). This term was proportional to the difference of the along-shelf surface pressure gradient at the detachment point and coastal boundary (Figure 15b).

The above analysis suggests a barotropic response of the current to an upwelling-favorable wind stress. When a uniform upwelling-favorable wind blows over the inner shelf, the near-surface current turns to the northeast direction in the region away from rivers. On the section connected to the river mouth, however, the Ekman effect speeds up the offshore river discharge current near the surface and hence causes the divergence and convergence zones on the northern and southern sides of a river, respectively (Figure 16). These convergence and divergence zones result in a northward decrease of the surface elevation in the along-shelf direction and thus increase offshore geostrophically dominantly transport in the surface mixed layer at the crest of the outer edge of the low-salinity front.

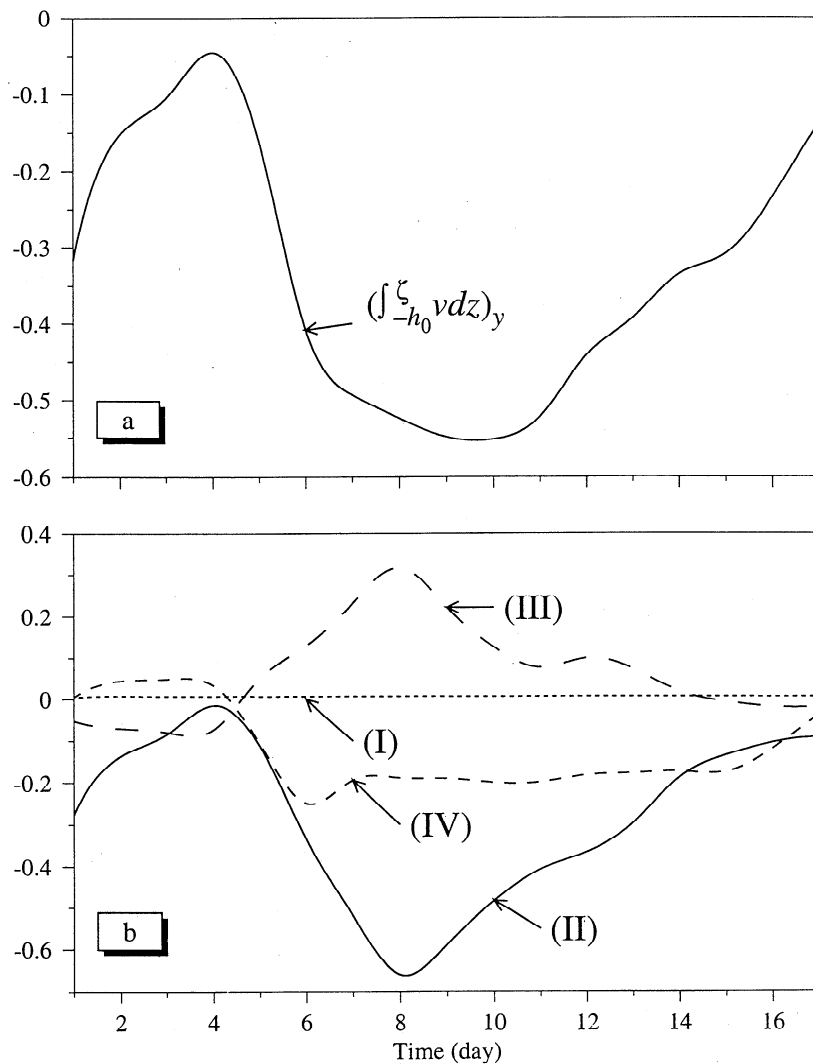


Figure 15. Time sequences of terms in (a) left-hand side and (b) right-hand side of equation (14). Unit is square meter per second.

The isolated low-salinity lenses form next owing to the along-shelf advection and vertical diffusion. This process can be clearly viewed from the temporal variation of the salt balance averaged over the upper 5 m from the surface at the outer edge of the front (Figure 17a). A significant decrease of the near-surface salinity occurred 4 days after the wind blew, which was caused by a strong offshore advection against the vertical salt diffusion. The near-surface salinity began to increase on the 9th model day after the wind blew owing to the positive vertical salt diffusion. An isolated low-salinity lens formed at the outer edge of the front 3 days later, at the time that the near-surface water was replaced by the relatively high salinity water advected from the upstream trough to the crest and diffused from the deeper region. Upwelling (the vertical salt advection) seemed to offer no significant contribution to the formation of the isolated lens in this case.

A quantitative estimation of the relative contributions of upwelling is given next. Figure 17b shows the temporal evolution of vertical velocity at the detached

point on section 2, 60 km off the coast. An upwelling developed during the 4th and 12th model days, which was characterized by two maxima: one was at 12 m on the 5th model day, and another was at 6 m on the 10th model day. The distribution of upwelling qualitatively implied that the isolated lens might be formed as an upwelling-induced outcropping tendency of haloclines. This implication, however, was not true in a quantitative sense. The halocline marked 34.6 was located at a depth of 8 m on the 10th model day; this halocline disappeared on the 12th day as a result of the outcropping of the 34.6 contour to the surface. If the outcrop of this salinity contour, which was found on the 12th model day, was caused only by upwelling, it would require a mean vertical velocity of

$$w = \frac{800}{2 \times 24 \times 3600} = 4.6 \times 10^{-3} \text{ (cm/s)},$$

which was twice as large as the mean vertical velocity shown in Figure 17b. In addition, the salinity contour of 34.8 outcropped the surface on the 14th day during a

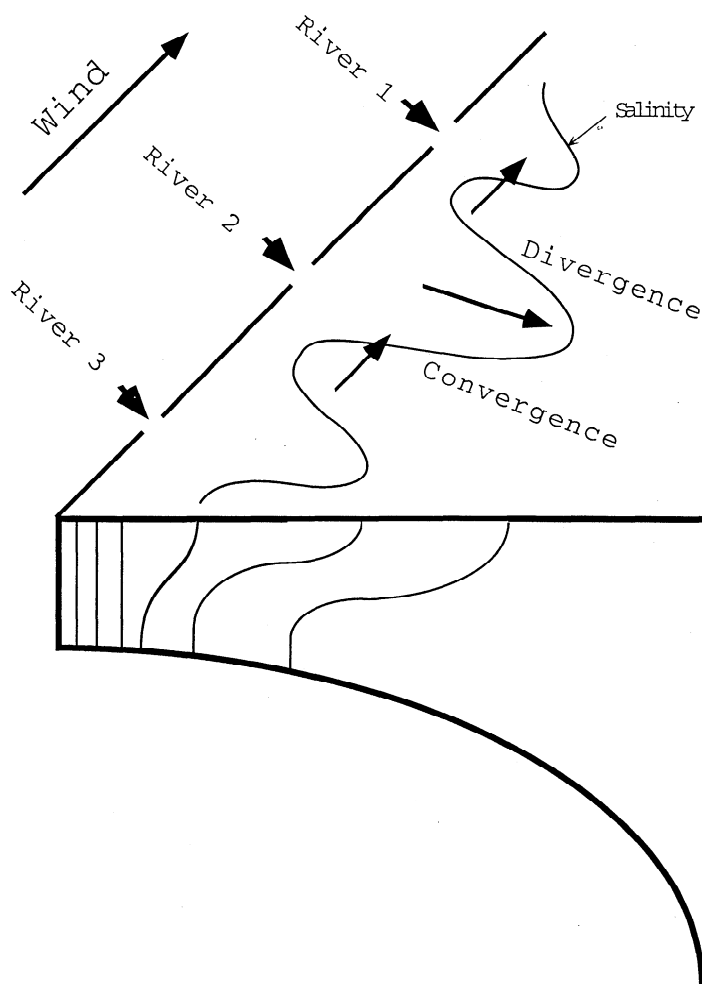


Figure 16. Schematic of the formation of divergence and convergence zones during an upwelling-favorable wind under the condition with multiple river discharges and tide.

period of downwelling. This implies that wind-induced upwelling was not a critical process that generated an isolated lens in this case. This analysis was consistent with our findings from the salt balance equation.

The relative importance of the northward along-shelf salt advection from the upstream trough to the crest also can be viewed quantitatively from a simple scaling analysis. The distance from the upstream trough to the crest was about 40 km on the 8th model day. The mean northward along-shelf velocity was slightly larger than 10 cm/s. If the water in the trough was advected northward at a speed of 10 cm/s, it would take about

$$T_{adv} = \frac{40 \times 1000}{0.1 \times 24 \times 3600} = 4.6 \text{ (days)}$$

This was the time at which the detachment of the isolated lens occurred.

Both salt balance and scaling estimations suggest that the formation of isolated low-salinity lenses can occur in two steps for the case with climatological conditions of river discharges and upwelling-favorable winds. At first, a nonuniform response of the inner shelf to an upwelling-favorable wind causes the along-shelf diver-

gence and convergence zones on the northern and southern parts of a river. These along-shelf divergence and convergence zones lead to a northward decrease of the surface elevation and hence cause a significant offshore low-salinity water transport on the cross-shelf section near larger discharge rivers. This process, as a result, forms a geometrically controlling wave-like shape of the salinity contours at the outer edge of the frontal zone. Isolated low-salinity lenses, then, form at the crest when water on the onshore side of the crest is displaced by relatively high salinity water advected from the upstream trough in the along-shelf direction and diffused vertically from the deeper region.

We also have conducted experiments excluding the mechanism of tidal forcing. The isolated low-salinity lenses disappear in these cases. This suggests that the interaction between tides, multiple river discharges, and winds over a 3-D shelf might play an important role in the detachment of low-salinity water across the front in the inner SAB shelf for the climatological condition of winds and river discharges. For a given upwelling-favorable wind of 5 m/s, tidal mixing tends to enhance the momentum transfer in the vertical, which

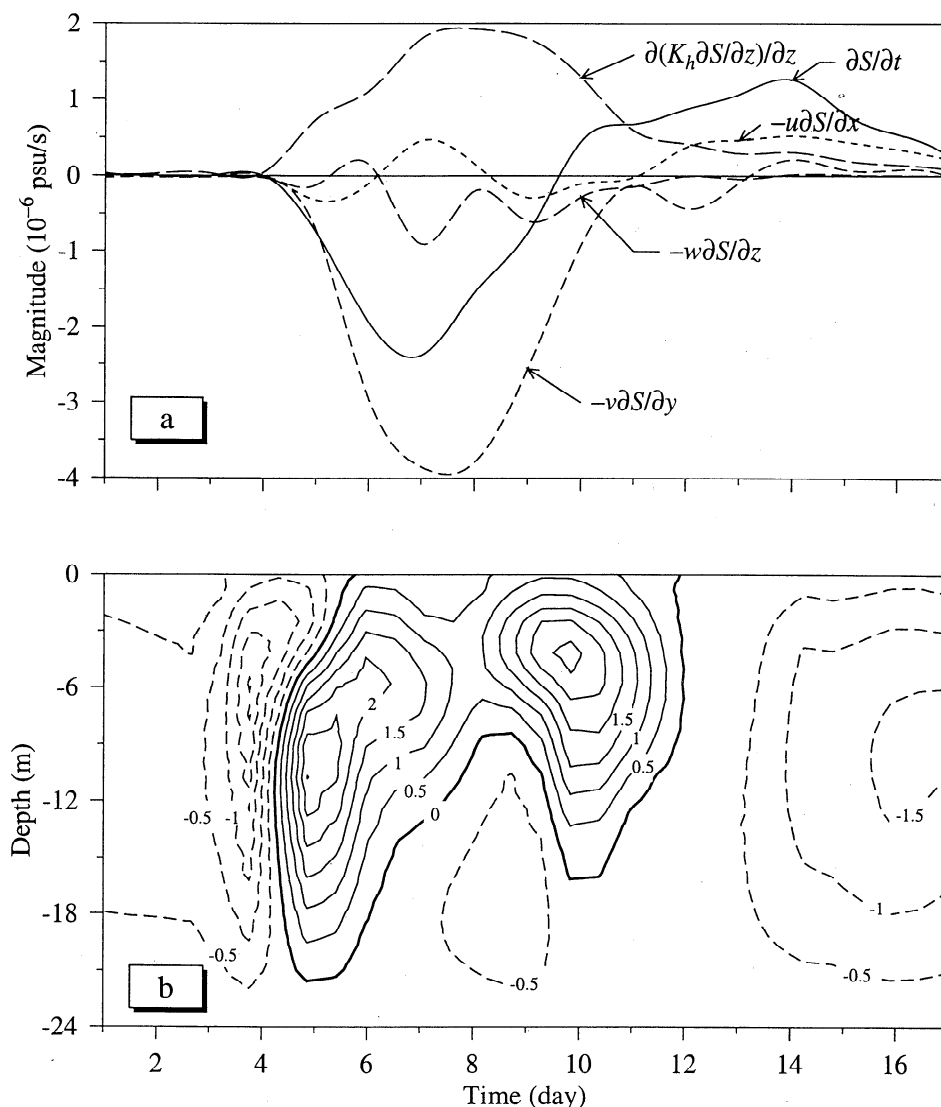


Figure 17. Time sequences of (a) each term in the salt equation (5) (in 10^{-6} psu/s) and (b) vertical velocity w (in 10^{-3} cm/s) at the detached point (60 km off the coast) from day 1 to day 14 after the wind blew.

produces a relatively weak wind-driven current at the surface. The combined current vector of river discharge and wind shifts significantly offshore near rivers, forming a wave-like shape of the low-salinity front (Figure 18: top). In the case without tide, however, the wind-induced momentum is mainly limited in a thin layer near the surface, which produces a much stronger along-shelf flow near the surface than that in the case with tide. The combined current vector of river discharges and wind is dominated by the wind-induced current, which is relatively uniform over the shelf (Figure 18, bottom). In this case, no wave-like shape of the salinity front forms!

It should be noted that our modeling experiments were conducted only for the climatological conditions of river discharges and wind. The episodic occurrence of isolated lenses in the inner shelf of the SAB is believed to be related to the direction and magnitude of

wind, river discharges, and background stratification. Kourafalou *et al.* [1996] used a 3-D model to simulate the spring 1984 hydrographic and current fields observed in the inner shelf of the SAB. They found that low-salinity water can be displaced offshore as an episodic event of detached isolated lenses under conditions of larger river discharge and upwelling-favorable winds. Our present studies show that the isolated lenses could form under a climatological condition with an average river discharge rate during an upwelling event. The occurrence of isolated lenses is caused by a complex nonlinear interaction between tides, multiple river discharges, and winds over a 3-D shelf. Are the physical processes found in this study still true for the cases with larger river discharges? Is it possible for the low-salinity front to become unstable when the river discharges increase? If so, what is the dynamic condition for baroclinic instability in the inner shelf of the SAB? These

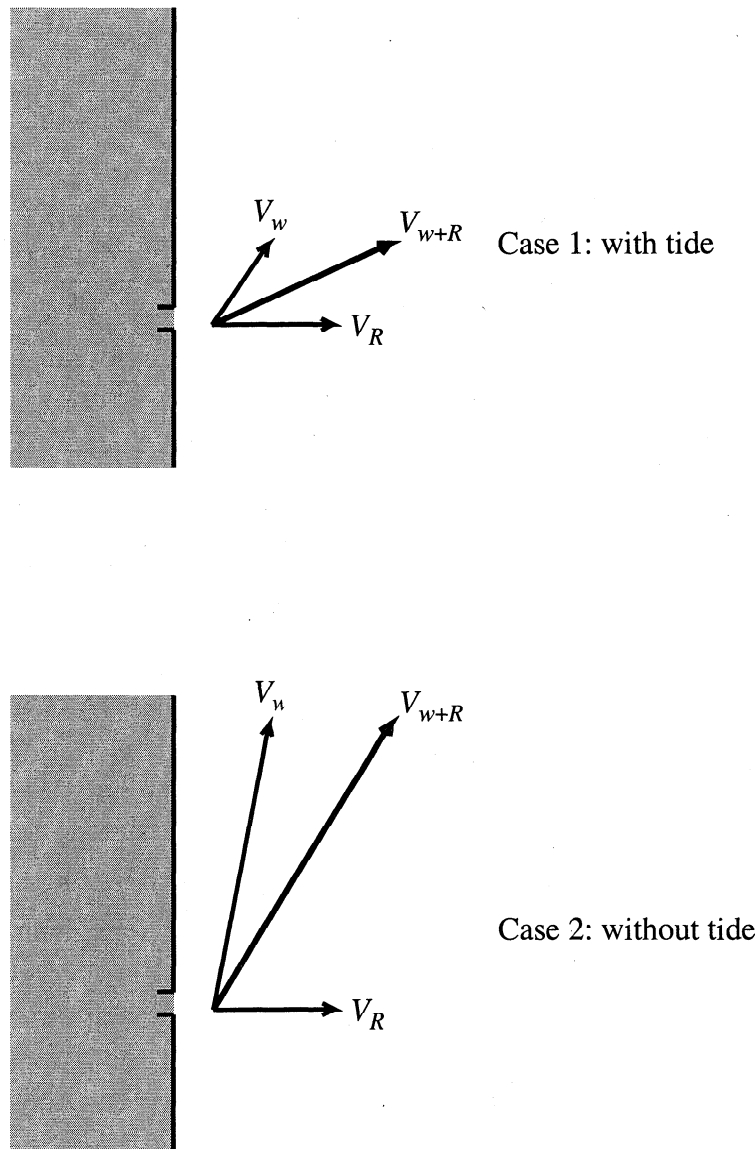


Figure 18. Schematics of combined current vectors of river discharge and wind for the cases with and without tides.

questions are being addressed by *C. Chen* (A modeling study of the cross-frontal water transport over the inner shelf of the South Atlantic Bight, submitted to *Journal of Physical Oceanography*, 1998) in a separate paper.

6. Conclusions

Physical processes that control the formation, evolution, and perturbation of the low-salinity front over the inner shelf of the South Atlantic Bight were examined using a fully three-dimensional primitive equation and turbulent closure model developed originally by *Blumberg and Mellor* [1987]. Process studies were conducted by running the model under different physical conditions with semidiurnal tides (M_2 , S_2 , and N_2), climatological means of multiple river discharges, and an upwelling-favorable wind. The model forced by semidiurnal tidal elevations (M_2 , S_2 , and N_2) has provided

a reasonable simulation of semidiurnal tides and the fortnightly and monthly tidal variations. The model-predicted amplitudes and phases are in good agreement with observational data available at Fernandina Beach, Florida, Savannah, Georgia, and Charleston, South Carolina.

The model predicts a topographically controlled along-shelf residual current near the coast for the case forced only by tides. Near the Savannah River, the residual current tends to flow northward parallel to shore, turn offshore, and then flow southward along the 10-m isobath. Interaction of tides with variable bottom topography off the Pee Dee and Cape Fear Rivers also generates weak, topographically controlled, anticyclonic residual eddies near the coast. The formation of topographically controlled residual currents is caused by barotropic tidal rectification over the variable bottom topography.

Spatial structure of buoyancy currents is significantly modified by stratified tidal rectification and tide-induced internal and bottom frictions. In the case forced only by river discharges, the model predicts a low-salinity plume that was trapped within the 20-m isobath over the inner shelf of the SAB. The low-salinity water within the plume is advected southward by a relatively uniform, along-shelf velocity. When the semidiurnal M_2 tide is added, the southward residual current is intensified significantly at the front and is reduced or reverses close to the coast. A weak velocity area forms in the downstream region of each river, which blocks the low-salinity water to form multiple "tongue-like" domes in the inner shelf, and slows or stalls the southward intrusion of the reduced-salinity zone near the coast. The model-predicted low-salinity frontal structure is qualitatively in good agreement with previous hydrographic observations and the spatial distribution of phytoplankton pigments observed from the satellite-produced coastal zone color scanner (CZCS) data.

For springtime climatological upwelling-favorable winds, the isolated low-salinity lenses could be episodically formed in two steps. At first, a geometrically controlling wave-like frontal shape develops at the outer edge of the frontal zone as a result of the interaction between tides, multiple river discharges, and upwelling-favorable wind. Then, the isolated low-salinity lenses form at the crest when water at the shoreward side of the crest is displaced by relatively high salinity water advected northward from the upstream trough and diffused vertically from the deeper region. Wind-induced upwelling does occur as a consequence of near-surface offshore Ekman transport, but it plays a secondary role in the formation of isolated lenses in the case with a climatological condition of multiple river discharges and upwelling-favorable winds.

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