

A three-dimensional simulation of coastal upwelling/downwelling off east coast of India

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Received 23 May 1997; revised 5 January 1998

Numerical experiments have been performed using the baroclinic 3-dimensional coastal ocean model. The model is used to simulate the coastal circulation off east coast of India. The driving mechanism consists, in general, of wind stress forcing representative of that applying during the premonsoon, monsoon and postmonsoon seasons. The simulations are in good agreement with observation on upwelling/downwelling off Visakhapatnam during different seasons.

Coastal upwelling is a term used to describe the coastal process whereby water from depths, generally not exceeding a few hundred meters, is brought upwards. It occurs within a distance from the coast of the order of the baroclinic Rossby radius of deformation¹ (about 10-50 km). As cold water from deeper depths which displaces the near water is generally rich in nutrients, upwelling zones are usually regions of high biological activity. The enhanced growth of phytoplankton and zooplankton in upwelling regions provide abundant food supply for the fish population. Consequently, the fisheries industry in the world depends heavily upon the regions of coastal upwelling. It is estimated that, in spite of constituting only 0.1 % of the area of the world's oceans, the regions of coastal upwelling account half of the world's fish production.

Generally speaking, coastal upwelling could be regarded as a result of one-sided horizontal divergence of the surface layers at the coast, the divergence itself being caused by an upwelling favourable wind which may be local or remote, or both. An upwelling-favourable wind may be defined as either an alongshore wind or a wind with an alongshore component with the coast to its left in the northern hemisphere and right in the southern hemisphere. The net transport is directed perpendicular to the right of the direction of the surface wind stress in the northern hemisphere, as shown in Fig. 1.

Regions of coastal upwelling, which are characterized by a reduction in the sea surface temperature and high biological activity, are identified mainly along the eastern boundaries of the oceans when prevailing winds associated with atmospheric circulation systems are upwelling-favourable. Other regions of coastal upwelling include part of the east coast of India where upwelling is caused by persistent southwesterlies from May to September².

Li³ has developed 3-dimensional continuously stratified model, with a non-dimensionalised horizontal co-ordinate in the offshore direction to yield increased resolution in the upwelling zone in addition to vertical sigma-co-ordinate (σ). Horizontal thermal advection at the boundaries was

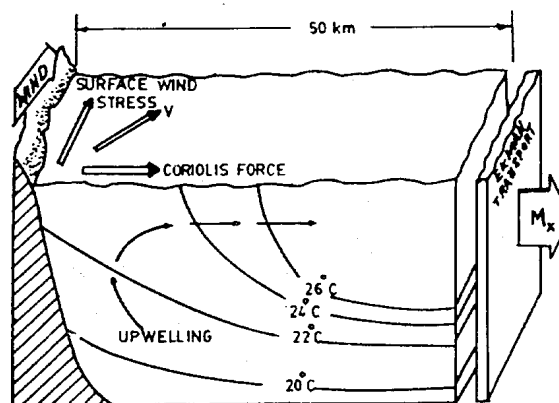


Fig. 1.—Dynamics of coastal upwelling in the northern hemisphere.

taken care of with upstream difference scheme and radiation boundary conditions are used at the open sea boundaries. With the model driven by the upwelling favourable wind Li³ studied the role played by coastally trapped waves in the dynamics of coastal upwelling.

Modelling studies of upwelling in the Bay of Bengal are very few and are also constrained by the limited observational studies. Rao *et al.*⁴ studied the upwelling process along the east coast of India. The numerical model used here is a baroclinic 3-dimensional model to study the ocean's response to wind stress forcing. In the present study, the simulations are carried out with this model to study the upwelling zones in different seasons, especially, in premonsoon, monsoon and postmonsoon period.

Materials and Methods

Model description

A local system of rectangular Cartesian coordinates Oxyz is used, with origin, O, located at the mean sea level and at the east coast of the southern open sea boundary. The coastal rigid boundary representing a part of the east coast of India is characterized by $x=0$ and the open-sea boundary corresponding to the $x=L_x$. The alongshore boundaries are fixed at $y=0$ and $y=L$. The displaced position of the sea-surface is given by $z=\zeta(x,y,t)$ and the position of sea-floor by $z=-h(x,y)$. Figure 2 shows schematically the model geometry and the coordinate system.

With a hydrostatically derived pressure gradient and Boussinesq approximation, the Reynolds averaged components of velocity (u, v, w) satisfy

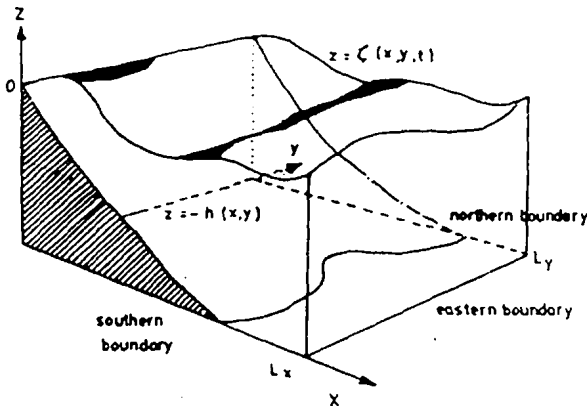


Fig. 2.—Model Geometry and coordinate system.

equations:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - fv = -g \frac{\partial \zeta}{\partial x} - \frac{g}{\rho_0} \int_0^{\zeta} \frac{\partial \rho}{\partial x} dz + \frac{\partial}{\partial z} \left(K_M \frac{\partial u}{\partial z} \right) \quad \dots (1)$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + fu = -g \frac{\partial \zeta}{\partial y} - \frac{g}{\rho_0} \int_0^{\zeta} \frac{\partial \rho}{\partial y} dz + \frac{\partial}{\partial z} \left(K_M \frac{\partial v}{\partial z} \right) \quad \dots (2)$$

where ρ is the local density of the water, ρ_0 is the reference density of the water and f is the Coriolis parameter.

The continuity equation of an incompressible fluid takes the form either

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad \dots (3)$$

The dynamical boundary conditions along the coastline leads to

$$u - vb'_1(y) = 0 \quad \text{at } x=0 \quad \dots (4)$$

where the prime denotes the differentiation with respect to y . Along the open-sea boundaries, we apply a modified version of the radiation type of boundary conditions⁴, which is of the form

$$\bar{u} - vb'_2(y) - \left(\frac{g}{h} \right)^{\frac{1}{2}} \zeta = 0 \quad \text{at } x=L_x \quad \dots (5)$$

$$\bar{v} - \left(\frac{g}{h} \right)^{\frac{1}{2}} \zeta = -2v_0 \quad \text{at } y=L_y \quad \dots (6)$$

$$\bar{v} + \left(\frac{g}{h} \right)^{\frac{1}{2}} \zeta = 0 \quad \text{at } y=0 \quad \dots (7)$$

where

$$(\bar{u}, \bar{v}) = \frac{1}{H} \int_{-h}^{\zeta} (u, v) dz \quad \dots (8)$$

are the depth-averaged components of the current and H is the total depth ($\zeta + h$). The v_0 is a velocity parameter, the value of which is chosen to simulate the fresh water discharge from the northern river system. In the case of premonsoon and postmonsoon experiments, the value of v_0 will be zero as they are

the lean periods fresh water discharge from the northern rivers.

At the sea-surface the internal Reynold stress is made equal to the applied surface wind stress. The sea-floor is characterized by no-slip condition leading to

$$u = v = w = 0 \quad \text{at } z = -h \quad (9)$$

The surface kinematical condition requires that

$$\frac{\partial \zeta}{\partial t} + u \frac{\partial \zeta}{\partial x} + v \frac{\partial \zeta}{\partial y} - w = 0 \quad \text{at } z = \zeta \quad \dots (10)$$

The equation of continuity gives a diagnostic form for w , leading to

$$w = - \int_{-h}^z \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dz \quad \dots (11)$$

and the same is written in the vertically integrated form as

$$\frac{\partial \zeta}{\partial t} + \frac{\partial}{\partial x} (H \bar{u}) + \frac{\partial}{\partial y} (H \bar{v}) = 0 \quad \dots (12)$$

The thermal energy, T , and salinity, S , satisfies an advection-diffusion equation of the form given by

$$\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} = \frac{\partial}{\partial z} \left(K_T \frac{\partial T}{\partial z} \right) \quad \dots (13)$$

and

$$\frac{\partial S}{\partial t} + u \frac{\partial S}{\partial x} + v \frac{\partial S}{\partial y} + w \frac{\partial S}{\partial z} = \frac{\partial}{\partial z} \left(K_S \frac{\partial S}{\partial z} \right) \quad \dots (14)$$

where K_T and K_S are the exchange coefficients for thermal energy and salinity respectively.

Here we prescribe that there is no vertical diffusive flux of thermal and salinity both at the sea-floor and sea-surface. Thus boundary conditions to accompany equations (13) and (14) leads to

$$\frac{\partial T}{\partial z} = \frac{\partial S}{\partial z} = 0 \quad \text{at } z = -h \text{ and } z = \zeta \quad \dots (15)$$

The coastal boundary is taken to be perfectly insulating with no heat transfer between the land and sea. At the open-sea boundaries the fluid flow determines the temperature and salinity across the boundaries.

The local density, ρ , is determined solely by the temperature and salinity of the water and is governed by a linear equation of state of the form

$$\rho = \rho_0 [1 - \alpha(T - T_0) + \delta(S - S_0)] \quad \dots (16)$$

The equation of turbulence energy density is written as

$$\begin{aligned} \frac{\partial E}{\partial t} + u \frac{\partial E}{\partial x} + v \frac{\partial E}{\partial y} + w \frac{\partial E}{\partial z} = \\ K_M \left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right] + \frac{\partial}{\partial z} \left[K_E \frac{\partial E}{\partial z} \right] \\ + \frac{g}{\rho_0} K_D \frac{\partial \rho}{\partial z} - \varepsilon \end{aligned} \quad \dots (17)$$

The first term on the right hand side of Eq. (17) represents the shear production of turbulence energy. The second term gives the redistribution of the turbulence energy by turbulence itself. The third term shows the suppressive effect on the production of the turbulence by a stable vertical density stratification and the last term represents the dissipation of the turbulence energy.

Boundary conditions to accompany Eq. (17) state that there is no vertical exchange of turbulence energy across either the sea-floor or the sea-surface leading to

$$\frac{\partial E}{\partial z} = 0 \quad \text{at } z = -h \text{ and } z = \zeta \quad \dots (18)$$

The representation of the exchange coefficients⁴ are given by,

$$K_M = c^{\frac{1}{4}} l E^{\frac{1}{2}} + N \quad \dots (19a)$$

$$K_T = K_S = K_E = c^{\frac{1}{4}} l E^{\frac{1}{2}} \quad \dots (19b)$$

The rate of dissipation of the turbulence energy, ε , in equation (17) and the turbulence length scale, l , in equation (19) are parameterized respectively as

$$\varepsilon = \frac{c^{\frac{3}{4}} E^{\frac{3}{2}}}{l} \quad \dots (20)$$

and

$$l = \frac{1}{\frac{1}{\lambda_1} + \frac{1}{\lambda_2}} \quad \dots (21)$$

where λ_1 and λ_2 have the form

$$\lambda_1 = \kappa(z_{os} + z_1) \quad \lambda_2 = \kappa(z_{ob} + z_2) \quad \dots (22)$$

where z_1 and z_2 are the vertical distances from the sea-surface and the sea-floor respectively. κ is Von Karman's constant, z_{ob} the roughness of the sea-floor, z_{os} a parameter related to the roughness of the sea-surface. The numerical solution of these

equations is obtained by integration ahead in time with the prescribed wind-stress forcing given for a particular month.

Numerical experiments

Numerical experiments have been performed by using the analysis area covering the coastal stretch shown in Fig. 3. The north-south extent of the area is about 800 km. We prescribe the wind-stress forcing in a limited area of approximately 450 km in the middle of the analysis area. The reason for prescribing the unforced regions in the south and north of the analysis area is well described in Rao *et al.*⁴. The breadth of the analysis area (the distance from the coastline to the eastern open-sea boundary) is constant and equal to 100 km. There are 20x23 grid-points in a horizontal computational plane and 36 computational levels through the depth. Each of the following experiments are performed by applying the wind-stress forcing for premonsoon (May), monsoon (July) and postmonsoon (November) period. An integration is carried out starting from a state of rest with an initial density field and continued for a period of 4 days to evolve vertical current and thermal structure off Visakhapatnam.

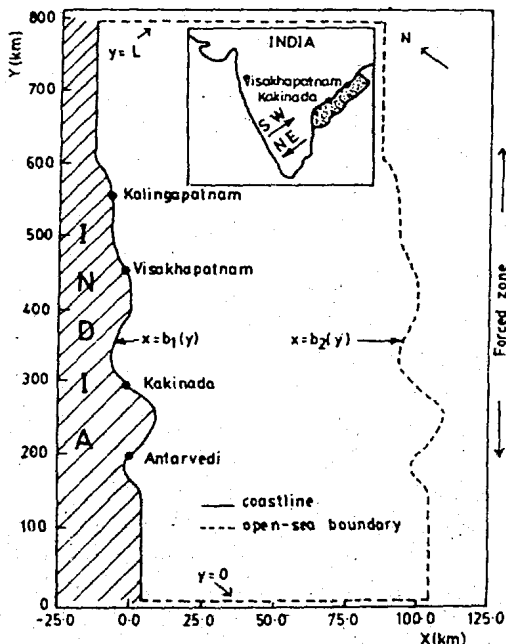


Fig. 3—Plan of the analysis area.

Results and Discussion

In the first experiment, the model is integrated with the mean monthly wind-stress forcing corresponding to May. In this case, the value of v_0 is taken as zero as there is very little fresh water flow from the northern river system. The southerly winds during the month are known to be quite favourable to coastal upwelling. It is consistent with the observations off Visakhapatnam reported by Rao & Rao⁵. The computed vertical thermal field off Visakhapatnam is shown in Fig. 4. The upwarping of the isotherms at all depths is clearly evident and also observed that there is a significant reduction of temperature of about 2°C at the sea surface. It is noticed that the 27°C isotherm which outcrops at the sea-surface shifts offshore by about 50 km. The current vectors in a vertical cross-shelf section corresponding to the thermal field off Visakhapatnam is shown in Fig. 5. Although the flow direction is not clearly seen in the figure, it is found that there is a weak offshore directed flow in

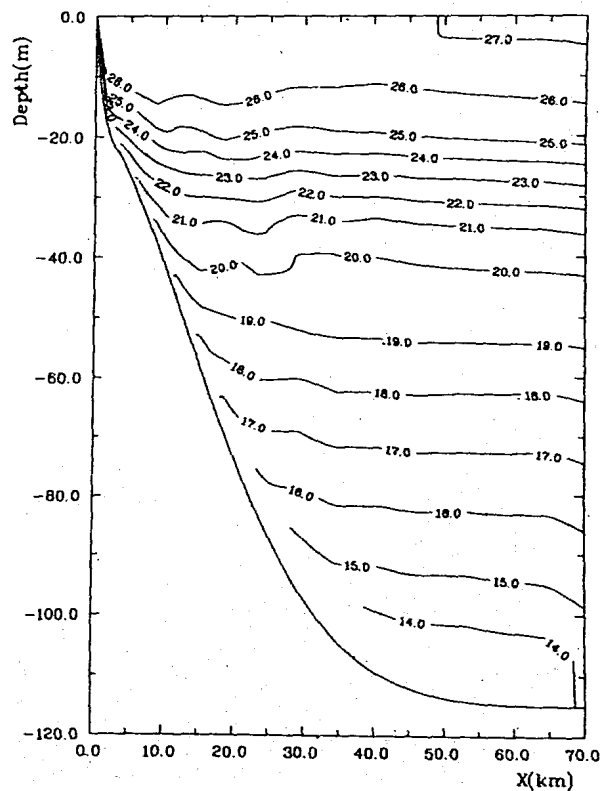


Fig. 4—The computed vertical thermal structure off Visakhapatnam for May

the surface layers with maximum speed of 1.0 cm s^{-1} . The existence of onshore flow both in the bottom Ekman layer and within the interior of the ocean at depths below the sea-surface greater than 20m may be seen. Maximum onshore current speeds are about 11.0 cm s^{-1} . The unique feature of the present study is the predicted detail of the onshore flow at the bottom Ekman layer and is only produced because of a full resolution of near bottom processes. This implies the upwelling is dominant along the coast which is consistent with the earlier observations off Visakapatnam reported by Rao & Rao⁵.

In the next experiment, the model is forced with the July wind stress forcing. During this month, a heavy fresh water flow from the northern river system is expected and as a result of this the coastal circulation will be changed⁶. In order to incorporate the freshwater discharge, a southward transport of water across the northern open-boundary is applied by taking a positive value for v_0 . The value of v_0 determines the amount of freshwater discharge which is pumping into the model. In Fig. 6, the thermal structure off Visakapatnam is shown. It is clearly seen from Fig. 7, which shows the vertical

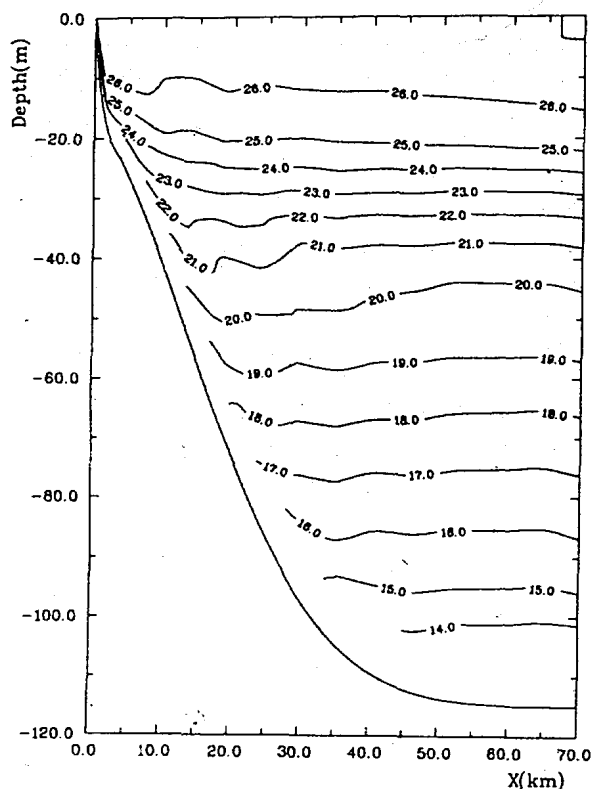


Fig. 6—The computed vertical thermal structure off Visakapatnam for July

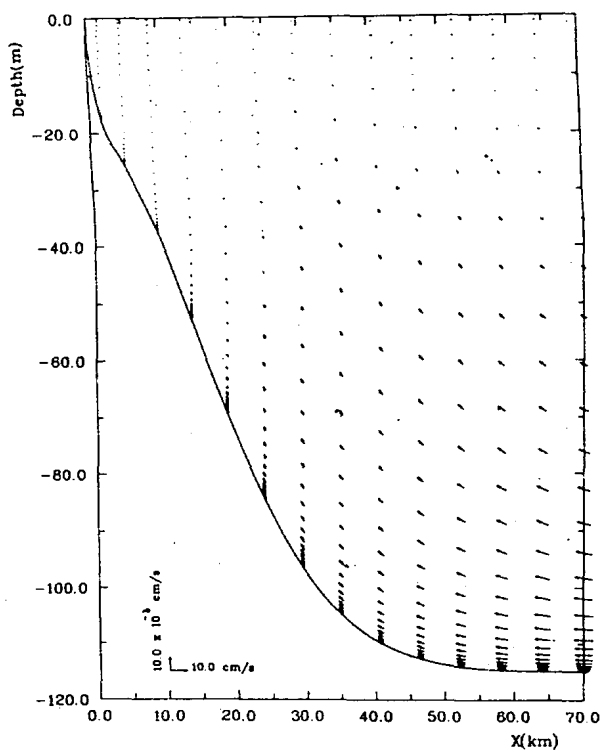


Fig. 5—The computed vertical current structure off Visakapatnam for May

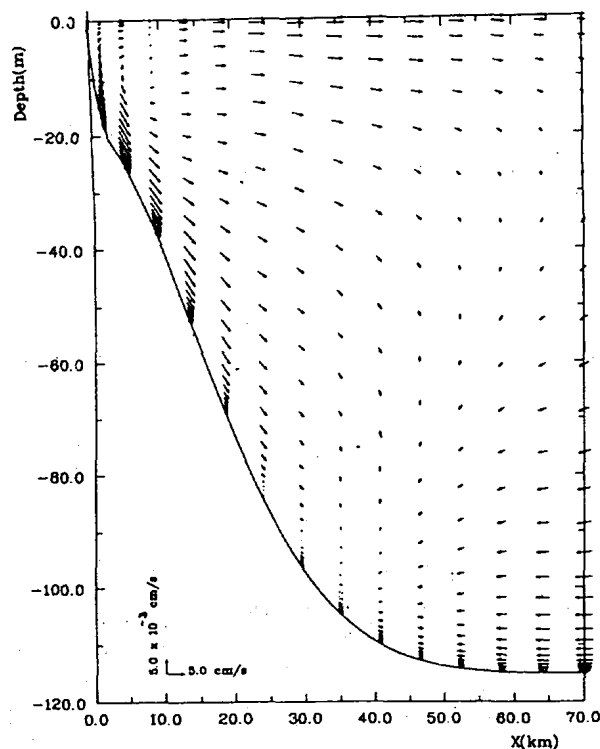


Fig. 7—The computed vertical current structure off Visakapatnam for July

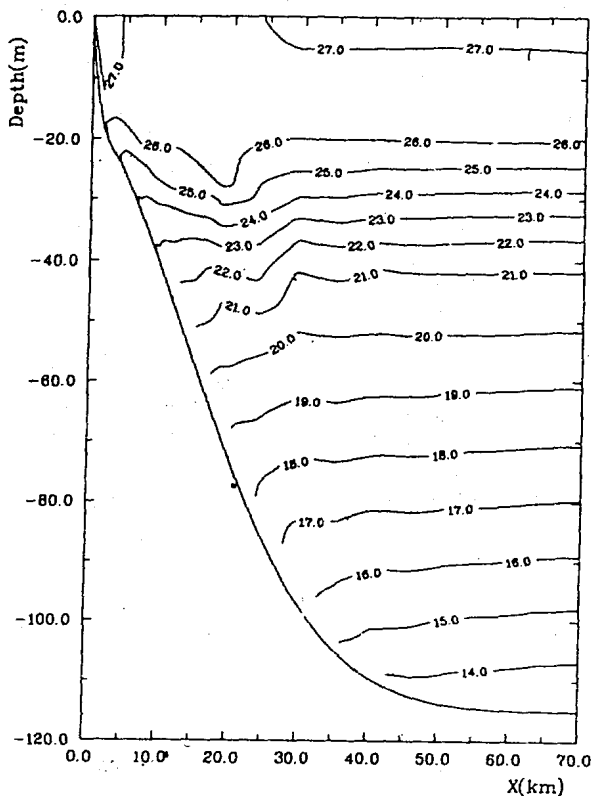


Fig. 8—The computed vertical thermal structure off Visakhapatnam for November

current structure that the coastal upwelling in May (Fig. 5) is replaced by downwelling with maximum sinking speed of $8.5 \times 10^{-3} \text{ cm s}^{-1}$. This experiment suggests that the coastal upwelling off Visakhapatnam in the premonsoon month is suppressed by the northern freshwater discharge and is, instead, replaced by local sinking during July. This predicted suppression of upwelling in the month of July is consistent with the reported observations⁶. In the final experiment, the November mean monthly winds are considered for integrating the model. Here, v_0 is set equal to zero as there is very little freshwater flow from the north. During this season, the winds are predominantly northeasterly. This suggests that no upwelling process is taking place during the period as the winds are not favourable. The predicted core of warm water in the near coastal region is consistent with the observations^{7,8}. From Fig. 8, the development of coastal upwelling is apparent in the marked downwarping of the isotherms during November from the initial field. This feature is more evident from the corresponding vertical current

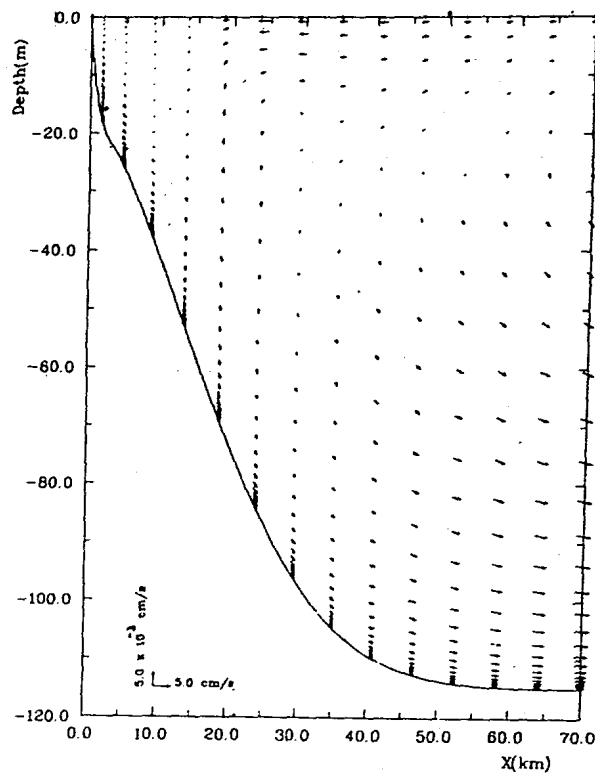


Fig. 9—The computed vertical current structure off Visakhapatnam for November

structure (Fig. 9). The maximum downwelling is about $3.4 \times 10^{-3} \text{ cm s}^{-1}$.

It is concluded that the simulations are compared with the observations⁵⁻⁸ and found that the model could simulate successfully the upwelling processes. The model also can be used to identify the upwelling zones along the coast and can be compared with the satellite SST observations for validation purpose.

References

- 1 Yoshida K, *Rec Oceanogr Wks Japan, New Ser*, 2 (1955) 8.
- 2 Lafond E C, *Proc Indian Acad Sci, Section B*, 46 (1957) 1.
- 3 Li Z, *A numerical study of basic coastal upwelling processes*, Ph.D. thesis, University of Reading, UK, 1990.
- 4 Rao A D, Sinha P C, Dube S K & Chamarthi S, *Proc Indian Acad Sci (Earth Planet Sci)*, 102 (1993) 465.
- 5 Rao T V N & Rao B P, *Indian J Mar Sci*, 18 (1989) 46.
- 6 Gopala Krishna V V & Sastry J S, *Indian J Mar Sci*, 14 (1985) 62.
- 7 Varadachari V V R, Murty C S & Das P K, *Bull Natl Inst Sci India*, 38(1968) 301.
- 8 Rao D P, *A comparative study of some physical processes governing the potential productivity of the Bay of Bengal and Arabian Sea*, Ph.D. thesis, Andhra University, India, 1977.