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Seasonal variation of residual flow in Suyoung Bay, Korea

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Abstract: In order to investigate the seasonal variation of residual flow in Suyoung Bay, Korea, we develop a robust diagnostic numerical model. The seasonal variation of residual flow in the upper layer at the inner bay is under the influence of the density-driven current by the fresh water supply from Suyoung River, that is, the residual flow in August shows the strong southward flow when the inflow of fresh water is large, but the residual flow is weak when the effect of fresh water supply is small in February. The residual flow pattern in the upper layer was also affected by the wind variation; the southwesterly wind in May and February generates the eastward flow, the northeasterly wind in August and the northwesterly wind in November generate the south-westward and southward flow, respectively. The characteristic of the flow pattern in the middle layer is the replenish flow, that is, the flow pattern in the middle layer is inflow when the flow pattern in the upper layer is outflow.

1. Introduction

Suyoung Bay is situated at the coastal area of Pusan, Korea. The oceanic condition of Suyoung Bay is very complicated because of the effects of the fresh water supply from Suyoung River and the offshore water mass from Korea–Tsushima Strait (Fig. 1).

Until now, many researchers have studied about the oceanic condition of Suyoung Bay. Particulary, the chemical study fo polluted material was done by WON and LEE (1979) and WON *et al.* (1979). Physical research by KIM and HAN (1982) was carried out on the diffusion phenomena in Suyoung Bay. KIM *et al.* (1991) studied the seasonal variation of oceanic condition in Suyoung Bay. However, the previous studies are only on the flow and diffusion at short period or the general characteristics of oceanic conditions.

There has been no study on the seasonal variation of residual flow which plays a very important role in the material transport during a long perid. The major components of the residual flow in the coastal sea are the tide-induced residual current, the wind-driven current and the density-driven current. In order to investigate the seasonal variation of residual flow in Suyoung Bay, we have developed a three-dimensional robust diagnostic numerical model which includes the effects of tide, wind and buoyancy. One of advantageous points of diagnostic numerical model is that three-dimensional velocity field is obtained quantitatively at each grid point in the model (e.g. BLUMBERG and MELLOR, 1987). The previous studies of circulation with use of a robust diagnostic numerical model are as follows for example; the diagnostic calculation of circulations and water mass movements in the deep Pacific Ocean (FUJIO and IMASATO, 1991) and the seasonal variation of circulations in the East China Sea and the Yellow Sea (YANAGI and TAKAHASHI, 1993).

In this study, we try to make clear the seasonal variation of residual flow in Suyoung Bay with the use of three dimensional robust diagnostic numerical model which includes the tide-induced residual current, the wind-driven current and the density-driven current.

2. Field observation

The horizontal and vertical distributions of water temperature, salinity and density were

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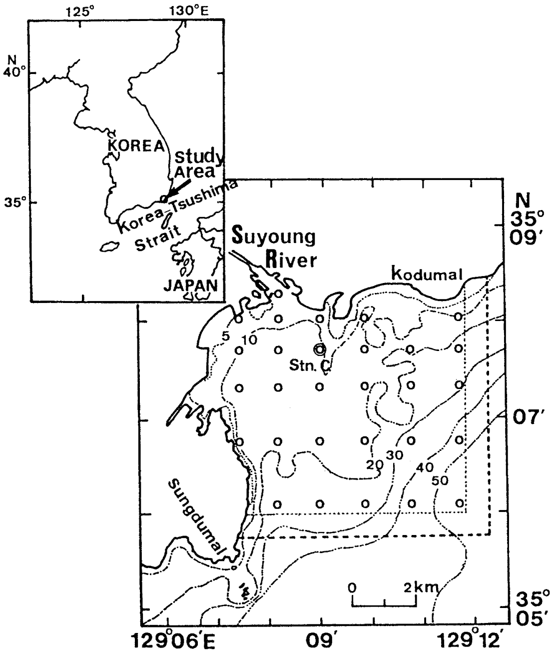

Fig. 1. Location of observation points in Suyoung Bay. The thick dotted line and the thin dotted line represent the open boundary of the numerical model and the observation area, respectively. Stn. C is the current observation station. Numbers show the depth in meter.

observed at 29 stations shown in Fig. 1 every month from May, 1989 to April, 1990 (KIM *et al*., 1991). The observed area is divided horizontally into 200mX200m mesh and vertically into three layers (0–5m, 5m–20m and 20m–bottom) and water temperature, salinity at every mesh are objectively interpolated from observed data with the use of a hyperbolic function (YANAGI and IGAWA, 1992). The upper layer (0–5m) expresses the mixed layer above the seasonal pycnocline.

The horizontal distributions of water temperature, salinity and density in spring (15 May, 1989), summer (17 August, 1989), autumn (25 November, 1989) and winter (27 February, 1990) are shown in Fig. 2(a) to 2(d), respectively. In spring, water temperature is high at the inner and outer bay, and salinity is low at the mouth of Suyoung River but high at the outer bay. Density distribution is affected by those of water temperature and salinity. In summer, water temperature at the inner bay is higher than that at the outer bay and salinity at the inner bay is lower than that at the outer bay. Density distribution is manily dependent on the salinity distribution. The characteristic of the distributions in autumn is that the differences of water temperature, salinity and density between the inner bay and the outer bay are generally small. In winter, water temperature at the inner bay is lower than that at the outer bay. Salinity and density distributions in winter are similiar to those in spring.

Year-to-year variation in the precipitation at Pusan is shown in Fig. 3. The precipitations in these observations from May 1989 to February 1990 are near to the averaged value during 5 years. The wind speed is not so strong (about 3~4m s^{-1}) and its direction is variable through out the year as shown later in Fig. 4; that is, the monsoon wind does not prevail in Suyoung Bay.

3. Numerical model

The horizontal grid size and the vertical division of water column of this numerical model are the same ones which are used for the interpolation of observed data. Depths used for this model were taken from Korean hydrographic chart No. 201C. Using conventional notation, the governing equations on the cartesian coodinate are as follows (YANAGI and TAKAHASHAI, 1993);

$$\frac{\partial \bar{u}}{\partial t} + (\bar{u}\cdot\nabla_h \bar{u}) + w\frac{\partial \bar{u}}{\partial z} + f\kappa \times \bar{u} = -\frac{1}{\rho_o}\nabla_h P + A_h \nabla_h^2 \bar{u} + A_v \frac{\partial^2 \bar{u}}{\partial z^2} + F_t, \qquad (1)$$

$$P = \rho_o g \xi + \int_z^{\xi} \frac{\rho_o - \rho}{\rho_o} g dz, \qquad (2)$$

$$\nabla_h \bar{u} + \frac{\partial w}{\partial z} = 0, \qquad (3)$$

$$\frac{\partial T}{\partial t} + (\bar{u}\cdot\nabla_h) T + w\frac{\partial T}{\partial z} = K_h \nabla_h^2 T + K_v \frac{\partial^2 T}{\partial z^2} + \Upsilon(T^* - T), \qquad (4)$$

$$\frac{\partial S}{\partial t} + (\bar{u}\cdot\nabla_h) S + w\frac{\partial S}{\partial z} = K_h \nabla_h^2 T + K_v \frac{\partial^2 S}{\partial z^2} + \Upsilon(S^* - S), \qquad (5)$$

where $\bar{u}$ is the horizontal velocity vector, w the upward velocity, f ($=2\omega \sin\phi$, ω is angular velocity of earth rotation and ϕ is latitude of

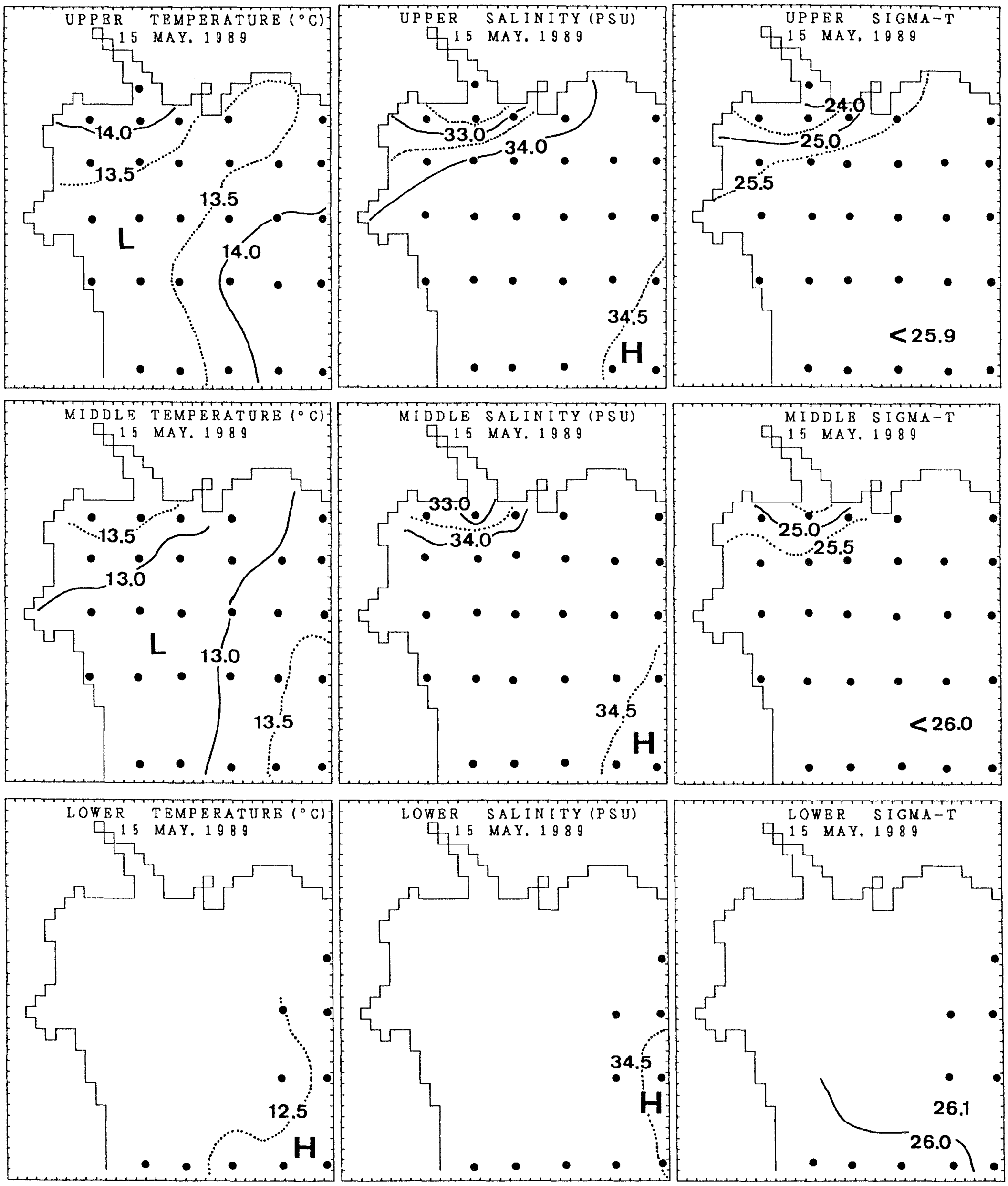

Fig. 2(a). Horizontal distributions of water temperature (left), salinity (central) and density (right) in three layers on 15 May, 1989. (●: observed stations)

35°05′ N) the Coriolis parameter, κ the vertical unit vector, ∇_h the horizontal gradient operator, t the time, P the pressure, ρ the density, ρ_0 the refference density, g the graviational acceleration ($=980$cm s^{-2}), ζ the sea level height above the mean sea surface, T water temperature, and S salinity. The density ρ is calculated from T and S with use of the usual nonlinear state equation. The Υ term in Eqs(4) and (5) is introduced by SARMIENTO and BRYAN (1982) to prevent calculated values T and S from deviating greatly from observed values T^* and S^*.

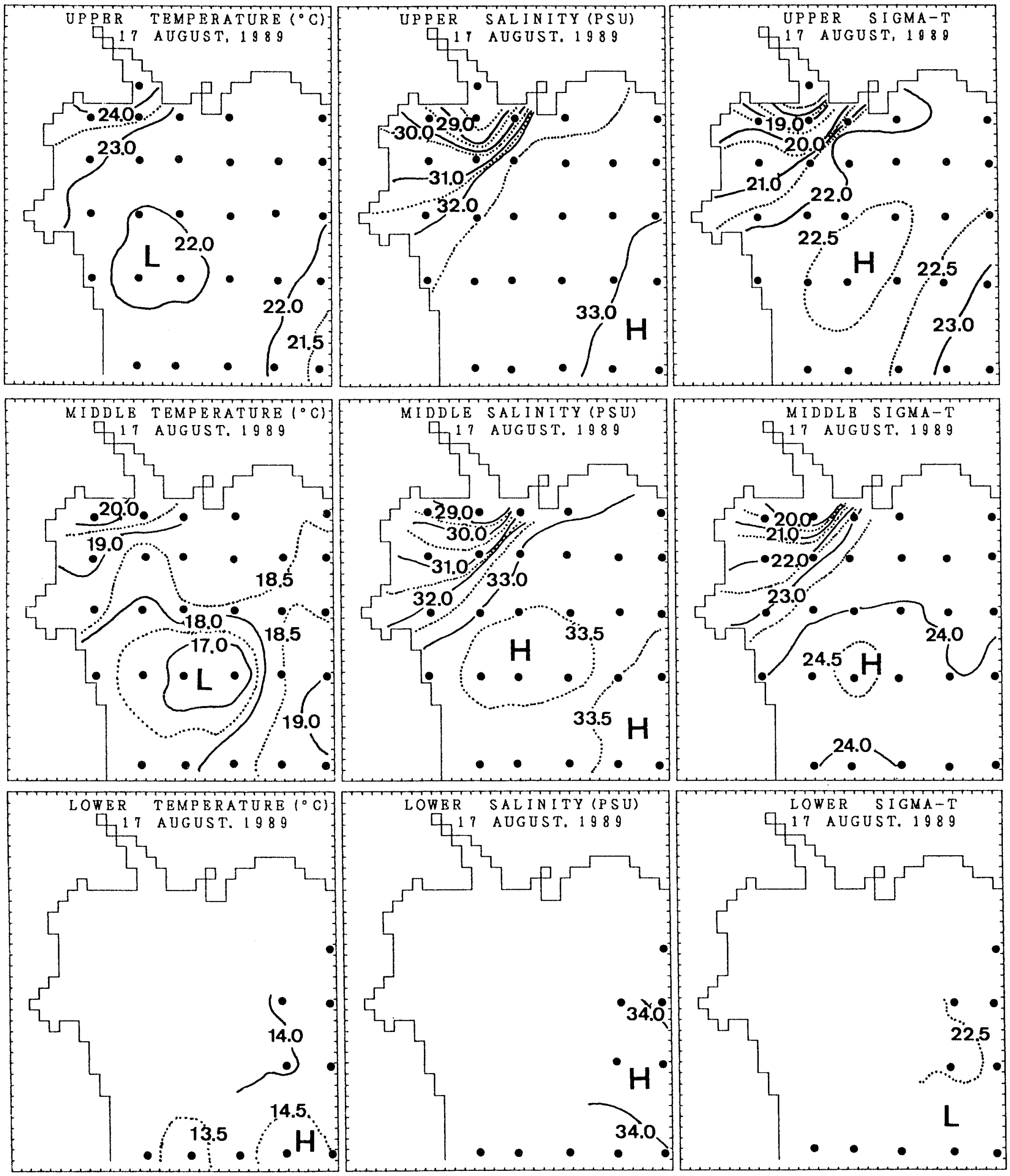

Fig. 2(b). Horizontal distributions of water temperature (left), salinity (central) and density (right) in three layers on 17 August, 1989. (●: observed stations)

Υ is $0.5/\Delta_t$ (Δ_t=30sec: time step of calculation) in this calculation. A_v and K_v are the vertical eddy viscosity and diffusivity, respectively. A_h and K_h are the horizontal eddy viscosity and diffusivity, respectively. The horizontal and vertical eddy viscosity and diffusivity are given as follows (YANAGI and YAMAMOTO, 1993);

$$A_h = \frac{0.3}{2\pi} \times U_{amp}^2 \times T_{M2}, \quad A_v = 10^{-5} \times A_h, \tag{6}$$

$$K_h = \frac{0.3}{2\pi} \times U_{amp}^2 \times T_{M2}, \quad K_v = 10^{-5} \times K_h, \tag{7}$$

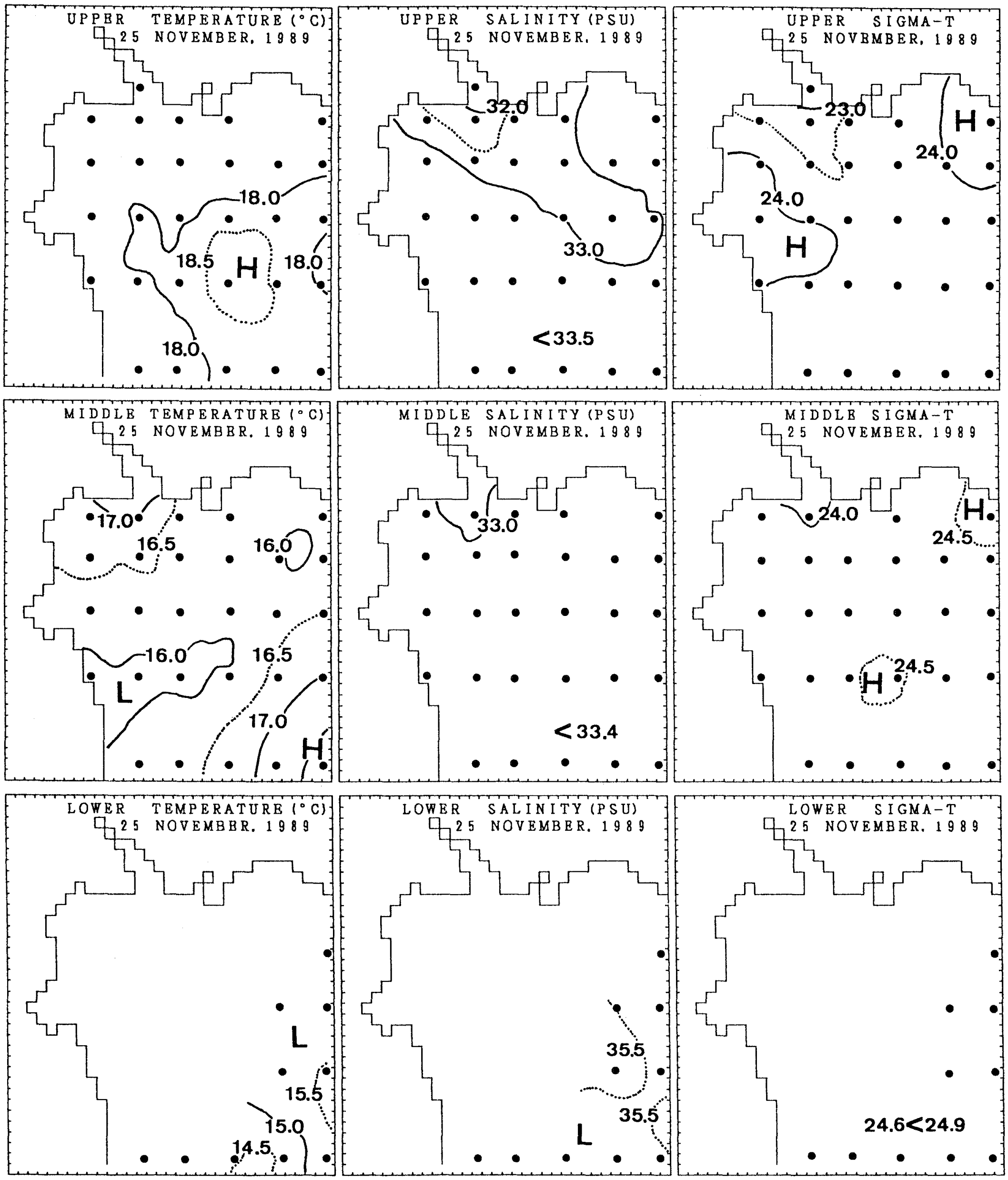

Fig. 2(c). Horizontal distributions of water temperature (left), salinity (central) and density (right) in three layers on 25 November, 1989. (●: observed stations)

where U_{amp} and T_{M2} denote the amplitude and period of M_2 tidal current, respectively. A_h and K_h range from 2.1×10^3 cm^2 s^{-1} to 3.4×10^6 cm^2 s^{-1} and A_v and K_v from 0.02 cm^2 s^{-1} to 34 cm^2 s^{-1} in Suyoung Bay. The horizontal forces due to tidal stress $F_t=(F_{tx},\ F_{ty})$ is calculated as follows (YANAGI and YAMAMOTO, 1993);

$$F_{tx}=-\overline{(U\frac{\partial U}{\partial x}+V\frac{\partial U}{\partial y})}, \tag{8}$$

$$F_{ty}=-\overline{(U\frac{\partial V}{\partial x}+V\frac{\partial V}{\partial y})}, \tag{9}$$

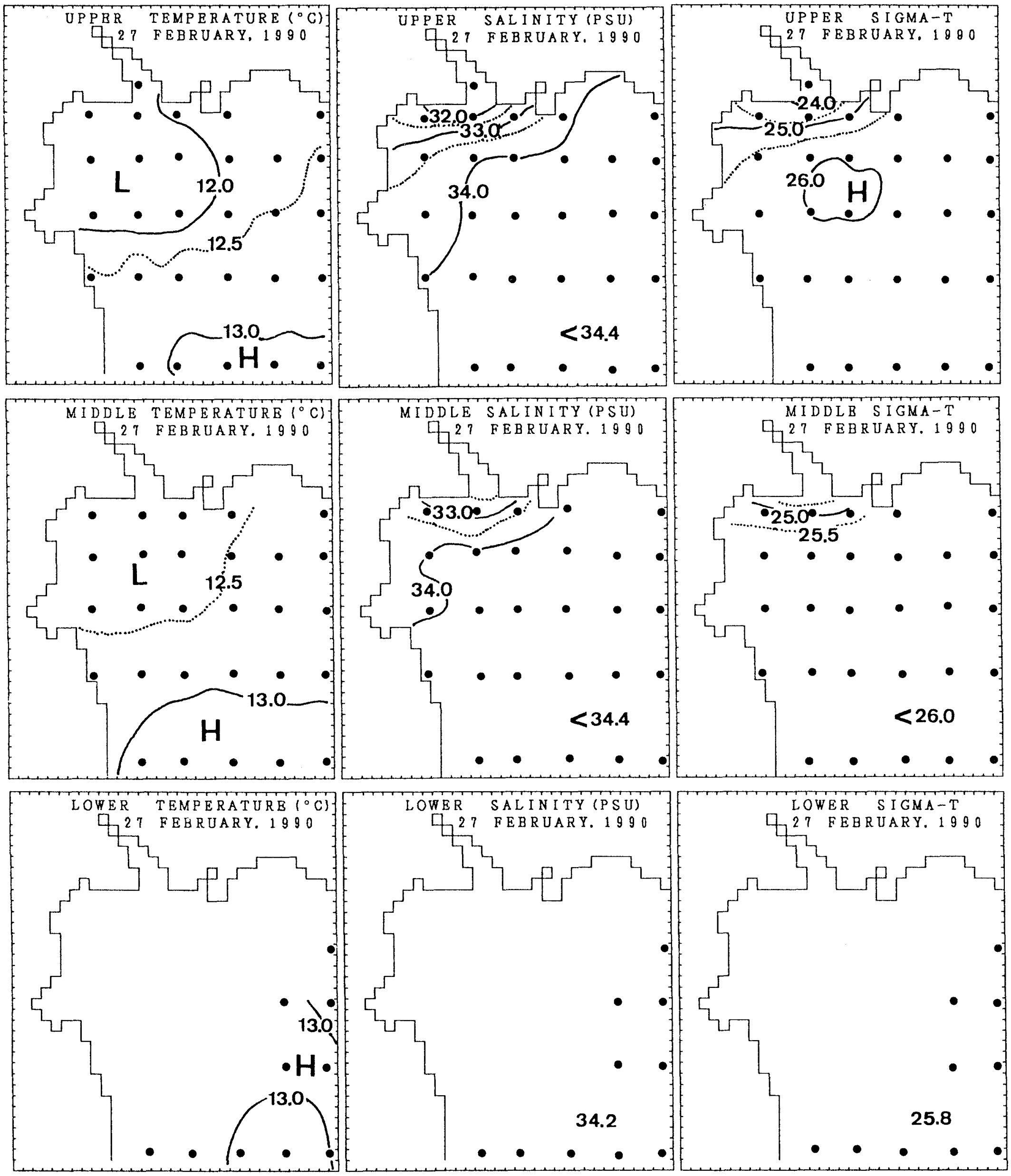

Fig. 2(d). Horizontal distributions of water temperature (left), salinity (central) and density (right) in three layers on 27 February, 1990. (●: observed stations)

where U and V are the x and y components of M_2 tidal current, respectively and the overbar denotes the average over one-tidal cycle. Tidal current in Suyoung Bay is already calculated and verified with the observational data (KIM and YANAGI, 1996).

The boundary condition for momentum is no-slip condition at the lateral wall. Along the open boundaries of numerical calculation, which is shown by the thick dotted line in Fig. 1, the normal velocity to the boundary is assumed to be zero. But the calculated results of

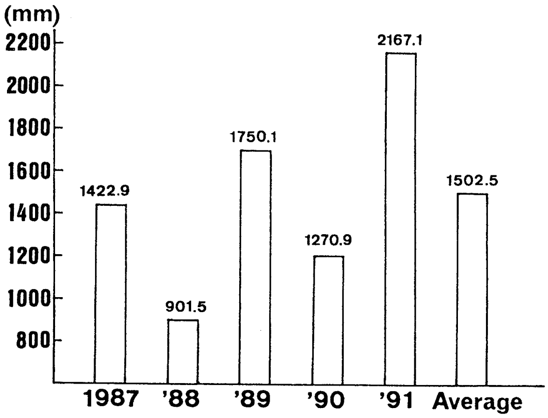

Fig. 3. Year-to-year variation in precipitation at Pusan, Korea from 1987 to 1991 and the averaged value.

residual flow in Suyoung Bay will be shown in the observation area which is the inner part of thin dotted line in Fig. 1. The bottom stress is given at the sea bottom as follows;

$$A_v \frac{\partial \bar{u}}{\partial z} = \Upsilon_b \, | \bar{u} | \, \bar{u} \,, \tag{10}$$

Here Υ_b (=0.0026) is the bottom drag coefficient.

The sea surface is assumed to be a free surface, and the sea surface momentum flux is given by

$$\rho_o A_v \frac{\partial \bar{u}}{\partial z} = \rho_a C_d \, | \bar{W} | \, \bar{W}, \tag{11}$$

Here ρ_a (=0.0012g cm^{-3}) is the air density, C_d (=0.0013) the sea surface drag coefficient and $\bar{W}$ the wind vector. The boundary condition for water temperature and salinity is a no-flux condition at the lateral wall, at the bottom,

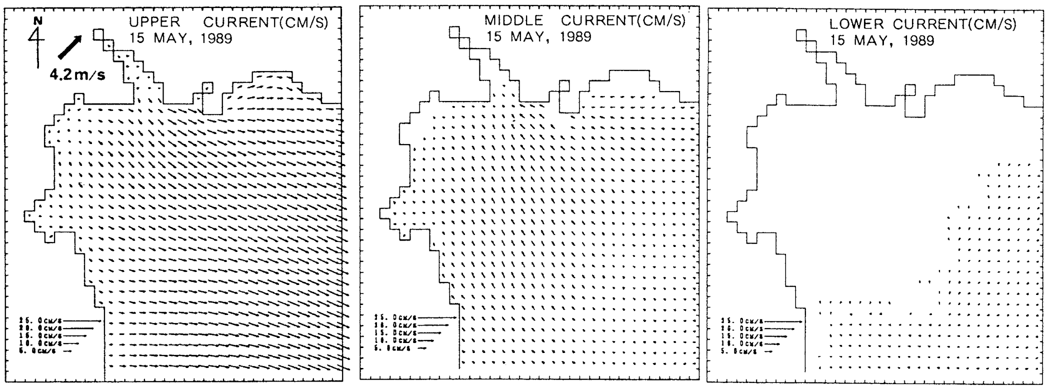

Fig. 4(a). Horizontal distribution of residual flow in three layers on 15 May, 1989. (thick vetor in the upper layer: wind speed and direction)

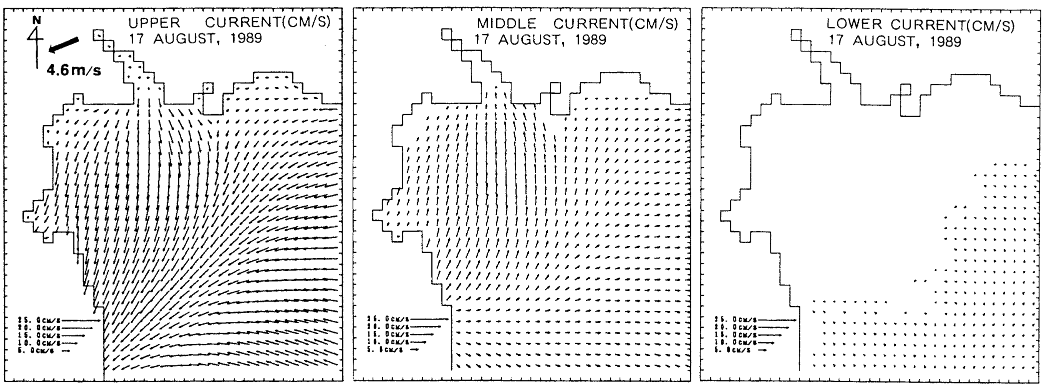

Fig. 4(b). Horizontal distribution of residual flow in three layers on 17 August, 1989. (thick vetor in the upper layer: wind speed and direction)

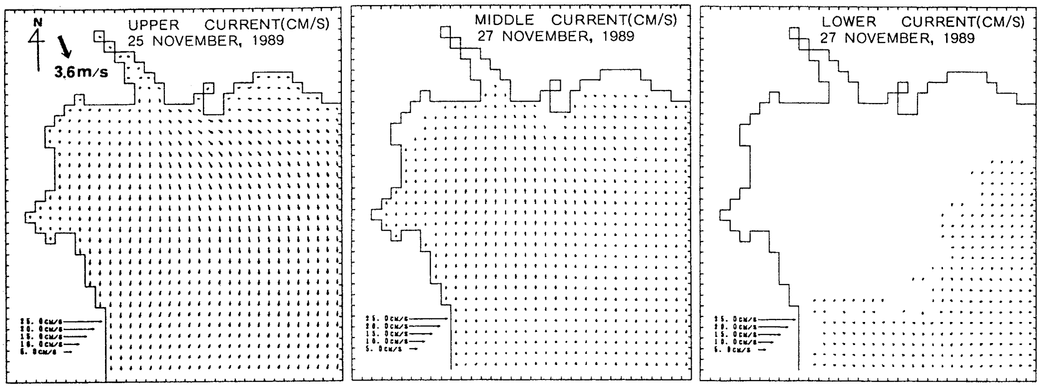

Fig. 4(c). Horizontal distribution of residual flow in three layers on 25 November, 1989. (thick vetor in the upper layer: wind speed and direction)

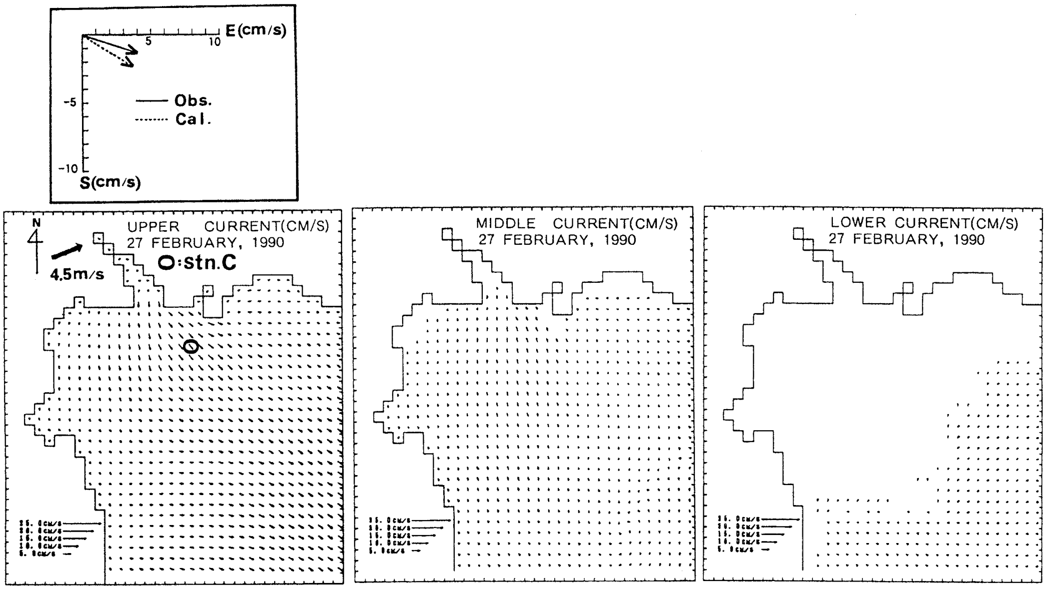

Fig. 4(d). Horizontal distribution of residual flow in three layers and the observed and calculated residual flows at Stn. C on 27 February, 1990. (thick vetor in the upper layer: wind speed and direction)

and at the sea surface. Along the open boundary, the horizontal gradient of water temperature and salinity are assumed to be zero. The central difference scheme is adopted for the advection term and the semi-implicit scheme is used for the calculation of sea surface elevation (BACKHAUS, 1983). Wind direction and speed are observed every hour at Pusan (Meteorological Observation Data of Korea, 1989, 1990). We used the daily averaged wind speed and direction on each observation day in this numerical model.

4. Results

The calculated flow patterns at three layers in spring (15 May, 1989) are shown in Fig. 4(a). The flow pattern in the upper layer is the eastward flow from the river mouth to the outer bay. In the middle layer, the flow pattern shows the northward current. In the lower layer, the flow pattern shows westward flow.

Figure 4(b) shows the flow patterns in three layers during summer (17 August, 1989). The flow pattern shows the south-westward flow from the inner bay to the outer bay in the upper layer. The flow pattern in the middle layer at the inner bay shows the strong northward flow but the flow pattern at the outer bay shows the eastward flow. The flow pattern in the lower layer is the northward flow.

The results in autumn (25 November, 1989) are shown in Fig. 4(c). The flow pattern in the upper layer shows the southward flow from the inner bay to the outer bay. In the middle layer, the flow pattern shows the inflow pattern, that is, the flow direction is northward. The flow in the lower layer is very weak. The flow patterns in the upper layer during winter (27 February, 1990) shown in Fig. 4 (d) are similar to those in spring, that is, the flow in the upper layer is east-southward from the inner bay to the outer bay.

In order to verify our calculated results, the comparision of calculated and observed flows is very important. The calculated and observed residual flows at Stn. C (see Fig. 1) are shown in Fig. 4 (d). The observation of current measuement was carried out 3m below the sea surface at Stn. C on 27 February (KIM *et al.*, 1991). The coincidence of calculated and observed flows is very good in speed and direction. Therefore, a robust diagnostic model used in this study is well verified with the direct observed result.

5. Discussion

To calculate the M_2 tide-induced residual flow, we used the two-dimensional numerical model of tidal current (KIM and YANAGI, 1996) because M_2 tidal current is the dominant current in this area from the report of the Pusan city of Korea (1984). The tide-induced residual flow at the inner bay is below 1 cm s^{-1} and it is very weak compared to the residual flow in the upper layer (5–10 cm s^{-1}) shown in Fig. 4. The tide-induced residual current is stable throughout the year because the tidal current is stable. Accordingly, the seasonal variation of residual flow shown in Fig. 4 is considered to be mainly generated by the density–driven current and /or the wind-driven current.

From Fig. 4, the residual flow shows the conspicious seasonal variation with the degree of density difference between the inner and outer bay in the upper layer. They are about 2.0 g cm^{-3} in May, 6.0 g cm^{-3} in August, 1.0 g cm^{-3} in November and 2.0 g cm^{-3} in February (see Fig. 2), respectively. The seasonal variation of residual flow speed in the upper layer at the inner bay is under the influence of the density-driven current by the density difference between the inner and outer bay in the upper layer. Such density differences are resulted from the fresh water supply from Suyoung River. The residual flow in August shows the strong southward flow when the inflow of fresh water supply is large, but the residual flow is weak when the effect of fresh water is small in November and February.

The pattern of residual flow in the upper layer of Suyoung Bay seems to be also affected by the wind condition. The seasonal variation of flow pattern by wind conditions is as follows; In May and February, the flow pattern in the upper layer is eastward by the southwesterly wind and that of August is south-westward by the east-north-easterly wind. In November, the north-west-northerly wind affected the southward flow. Therefore, the wind direction affected to the flow pattern in the upper layer to some degree. Such phenomena about the wind effect are well explained by the study on the variability of residual flow in Osaka Bay, Japan (YANAGI and TAKAHASHI, 1988).

The flow pattern in the middle layer generally shows the replenishment flow oppositing to that in the upper layer, that is, when the flow is outflow in the upper layer that is inflow in the middle layer.

The effect of the density-driven current is large by the inflow of fresh water from Suyoung River at the inner bay and the residual flow pattern in the upper layer is also affected by the wind-driven current. Therefore, when we consider about the residual flow fieled in Suyoung Bay, we have to take into account of not only the effect of tide but also the effects of wind and density distribution.

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