Volume and heat transports of the Kuroshio in the East China Sea in 1989

Yaochu YUAN*, Jilan SU* and Ziqin PAN*

Abstract: A modified inverse method is used to compute the Kuroshio in the East China Sea with hydrographic data and moored current meter records obtained during early summer and autumn in 1989. In this method the geostrophic assumption is not imposed. The vertical viscous term in the momentum equation as well as the vertical diffusion term in the conservative equation and the heat exchange at the surface are all considered. The computed volume transports at R. (PN) section are 35 and 31 × 10^6 m^3/s during early summer and autumn, respectively, in 1989. The heat transports are 2.50 and 2.29 × 10^19 W there during early summer and autumn, respectively, in 1989. About 1.29 × 10^19 W of heat is transferred into the ocean from the atmosphere over the region of computation during early summer of 1989 and about 2.80 × 10^19 W of heat from the ocean to the atmosphere during autumn of 1989. A counter-current is present in the deep layer during both periods. The total volume transport of the Taiwan Warm Current is about 1.5 × 10^6 m^3/s.

1. Introduction

Most of the previous studies on the volume transport of the Kuroshio in the East China Sea were based on the dynamic method (e.g. GUAN, 1982 and 1988; NISHIZAWA et al., 1982). The level of no motion is normally set at 7 MPa. The annual mean volume transport thus computed for the Kuroshio through the PN section in the East China Sea is usually around 20 × 10^6 m^3/s. (The position of the PN section is the same as section R. shown in Fig. 1(a)). Recently YUAN, ENDOH and ISHIKAWA (1991) and YUAN, SU and PAN (1991) applied the inverse method to compute the current structure and volume transport of the Kuroshio in the East China Sea. Their results showed that the conservation of mass is not satisfied by the dynamic method and approximately satisfied by the inverse method. In these studies the geostrophic relation was assumed, as is the normal practice for inverse method (WUNSCH, 1978). However, the geostrophic assumption is not always valid (YUAN et al., 1986).

In this paper, a modified inverse method is proposed to compute the current structure of the Kuroshio in the East China Sea. The hydrographic data and moored current meter records used were obtained in early summer and autumn of 1989.

2. Modified inverse method

Both the vertical friction term in the momentum equation and the vertical diffusion term in the density equation are kept in our modified inverse method. In addition, the heat flux through the sea surface is also considered. In the following we shall discuss the rationale and results.

1) The vertical friction term in momentum equation is kept because the East China Sea shelf is shallow. If we further take the rigid-lid assumption, the boundary conditions at the surface are then

\[ \rho_0 \Lambda z \frac{\partial u}{\partial z} = \tau_x, \quad \rho_0 \Lambda z \frac{\partial v}{\partial z} = \tau_y, \]

w = 0 \quad (1)

where the x-axis is parallel to hydrographic section, the y-axis normal to the section, and the z-axis upward. \( \tau_x, \tau_y \) is the wind stress vector, \( u \) and \( v \) are the horizontal velocity components, and \( w \) the vertical velocity component.

No-slip boundary condition is specified at the
bottom, i.e., at $z = -H(x, y)$

$$u = v = w = 0$$

(2)

For constant $A_z$ an approximate expression for the horizontal velocity can be found (see YUAN et al., 1986). For example, the $v$ component can be written approximately as

$$v = v_s + v_0 + v_b$$

(3)

$v_s$ is the surface Ekman velocity component subject to the wind stress $\tau$, $v_0$ is the geostrophic velocity, i.e.,

$$v_0 = b_i - \frac{\rho}{\rho_0} \int_{z_0}^{z} \frac{\partial \rho}{\partial x} \, dz$$

(4)

where $b_i$ is the value of $v_s$ at the reference level $z_0$ and is an unknown to be computed. $v_b$ is the bottom Ekman velocity components. If the pressure gradient is mainly in the along-section direction, i.e., $|\partial p/\partial x| > |\partial p/\partial y|$, then $v_b$ can be expressed simply as (YUAN et al., 1986):

$$v_b = -2 \left( \frac{\sin \alpha h}{\cos \alpha h + \sin \alpha h} \right) v_0$$

(5)

where

$$a = \left( \frac{f}{2A_b} \right)^{1/2}$$

2) SARKISYAN (1977) has shown that in either the density or salt equation, the convection terms are of dominant importance and the horizontal diffusion terms are smaller than the vertical diffusion term. If we keep all convection terms and the vertical diffusion term in the density and salt equation but drop the horizontal diffusion terms, then the following equations for the $j$ th layer of box $i$ can be obtained:

$$\sum A_i \rho (x, y, z) C(x, y, z) u (x, y, z) \delta_{i,j} dA_{i,j}$$

$$= C_i \rho \omega_j \delta_{i,-} A_{i,-} C_{i,-} \rho \omega_j \delta_{j,-} A_{j,-} + K_i \left[ \sum A_i \frac{\partial \rho}{\partial z} \, dA_i - \sum A_i \frac{\partial \rho}{\partial z} \, dA_{i,-} \right]$$

(6)

$K_i$ is the vertical diffusion coefficient, assumed to be constant, $\delta_{i,j} (\delta_{i,-}, \delta_{j,-})$ takes the value of 1 for flow into the box and -1 for flow out of the box. When $C(x, y, z) = 1$, Eqn. (6) represents the mass conservation, and when $C(x, y, z) = S(x, y, z)$. Eqn. (6) represents the salt conservation equation. $A_i$ is the lateral face for the $j$ th layer of box $i$, $A_{i,-}$ and $A_j$ are the projection of the upper and bottom faces of the $j$ th layer of box $i$ on the horizontal plane, respectively. In the standard inverse method, all terms on the left hand side of Eqn. (6), i.e. the vertical fluxes in the conservation equation, are neglected (e.g. WUNSCH, 1978). However, as we shall see later, the vertical fluxes due to the surface and bottom Ekman layer effect cannot be neglected in the deep layer.

From Eqns. (3)-(6), we can write the following matrix

$$Ab = \Gamma$$

(7)

where $b$ is $N \times 1$ matrices, and unknowns $(b_i, \omega_j)$, $A$ and $\Gamma$ can be computed from the hydrographic and wind data. 3) Considering that the heat flux $q$, at the sea surface is unknown, the following constraints are imposed, i.e.,

$$q_{e} \leq q_{s} \leq q_{o}$$

(8)

where $q_{e}$ and $q_{o}$ are, respectively, minimum and maximum statistical monthly average values in the survey area. We take the values of $q_{e}$ and $q_{o}$ as given by the Institutes of Oceanography and Geography of the Academia Sinica in 1977. $q_{s}$ is an unknown to be computed. Positive value of $q_{s}$ indicates heat transfer from the ocean to the atmosphere, and negative value of $q_{s}$ indicates the heat transfer from the atmosphere to the ocean.

Eqn. (7) is solved with the inequalities (8) imposed. This is a mathematical problem of quadratic programing. Details of the computation of the problem can be found in WUNSCH (1982). If Eqn. (7) is solved without the inequalities (8), it is a linear programing problem (e.g. FIADEIRO and VERONIS, 1982; YUAN et al., 1990).

On the choice of an optimum reference level, we use the FIADEIRO and VERONIS's method.
(1982). Each box is divided into $K$ layers. The
geostrophic velocity $\mathbf{v}$ relative to $z_i$ level is
given by

$$v = -\frac{\partial}{\partial z} \int_{z_a}^{z_i} \frac{\partial \rho}{\partial z} \, dz$$  \hspace{1cm} (9)

We define

$$T_{e}=\frac{1}{3} \int_{z_a}^{z_i} v \, dz \, dx, \quad (e=1,2,\ldots;M)$$  \hspace{1cm} (10)

and

$$T_i=\sum_{e=1}^{M-1} \sum_{j=1}^{K} T_{e}^j$$  \hspace{1cm} (11)

where $M$ is the number of boxes. A good choice for
a reference level $z_i$ is such that $T_i$ is the mini-
mum. If $z_a>H$(the water depth) at some stations,
the reference levels at these stations are
reset to be $H$.

The volume transport (VT) $Q_a$ is defined as

$$Q_a = \int_L dx \int_{-H}^{0} v \, dz$$  \hspace{1cm} (12)

where $L$ is the width of the section, Substituting
Eqn. (3) into Eqn. (12), we obtain

$$Q_a = Q_{a,\tau} + Q_{a,\sigma} + Q_{a,s}$$  \hspace{1cm} (13)

where

$$Q_{a,\tau} = \int_L dx \int_{-H}^{0} v \cdot \tau \, dz$$

$$Q_{a,\sigma} = \int_L dx \int_{-H}^{0} u \sigma \, dz$$

$$Q_{a,s} = \int_L dx \int_{-H}^{0} u_s \, dz$$  \hspace{1cm} (14)

$Q_{a,\sigma}$ is the geostrophic transport, $Q_{a,\tau}$, is the
surface Ekman transport due to the surface wind stress
and $Q_{a,s}$ is the bottom Ekman transport due to the bottom friction.

The heat transport $Q_i$ due to ocean currents
across a section $S$ is well approximated by

$$Q_i = \int_S \rho \, C_v \, T \, d\mathbf{s}$$  \hspace{1cm} (15)

Let

$$\tilde{v} = \tilde{\mathbf{v}} + \tilde{\mathbf{v}^\prime}$$

$$T = \tilde{T} + T^\prime$$

in which $\tilde{\mathbf{v}}$ and $\tilde{T}$ are the depth-averaged velocity
and temperature, respectively, $\mathbf{v}^\prime$ is the baro-
clinic velocity component and $T^\prime$ is the devi-
tions from the average temperature. Then Eqn. (2)
becomes

$$Q_i = \int_S \rho \, C_v \, \tilde{T} \, d\mathbf{s} + \int_S \rho \, C_v \, T^\prime \, d\mathbf{s}$$

The heat transport (HT) components can be
now divided into two parts, namely, the
barotropic and baroclinic components.

3. Data and parameters

In this study both the hydrographic data and
moored current meter data are used. During the
early summer cruise in 1989, a short-term (5
May–7 May) moored station $M_i$ was located on
section $S_i$ (Fig. 1.(a)) at 241 m water depth.
One current meter was set at 80 m depth and the
observed average velocity is 8.2 cm/s and 24°
clockwise from north. During the autumn cruise
in 1989, a moored station $M_i$ was located on section
$S_i$ (Fig. 1.(b)) at 100 m water depth. Two
current meters were set at 40 and 80 m depths,
and their observation period was from Oct. 11
to Oct. 22. The time-averages of the observed
velocities at 40 and 80 m depths were (14.6
cm/s, 31°) and (13.5 cm/s, 8°), respectively.
For this computation, only the observed velocity
at 80 m level is used as a known value of velocity
at the 80 m reference level in Eqn. (7).

Both the vertical diffusion coefficient $K_i$ and
vertical eddy coefficient $\Lambda_i$ are assumed to be
constant. We have experimented with different
values of these coefficients, namely, $A_i=50$ and
100 cm$^2$/s and $K_i=1$ and 10 cm$^2$/s.

The average wind direction and speed as on-
board R/V Shijian were 223° (SW) and
6.5 m/s during the early summer cruises of 1989
and 43.7° (NE) and 9.3 m/s during the autumn
cruise of 1989, respectively. Because of the lack of
accurate wind data, a steady uniform wind field
for each cruise with the above respective
average values is assumed.

To compare the effects of the vertical viscous
terms and the vertical diffusion terms we try
two values of vertical eddy coefficient $A_i$ (50
and 100 cm$^2$/s) and two values of vertical
diffusion coefficient $K_c$ (1 and 10 cm$^2$/s). The values of $q_i$ and $q_s$ are taken to be, respectively, $-1.7$ and $0.8 \times 10^8$ J/cm$^2$ d for the early summer cruise and 1.3 and $2.1 \times 10^8$ J/cm$^2$ d for the autumn cruise (see Institute of Oceanography and Geography, 1977). For comparison we have also carried out computations without imposing the inequality constraints.

4. Numerical experiments

Figures 1. (a), (b) show the bathymetry, hydrographic sections and computation boxes for the early summer and autumn, respectively, in 1989. The computation points are at the mid-points between neighboring hydrographic stations. All boundary sections of the computation boxes are divided into layers according to isopycnal $\sigma$ values of 24, 27, 30 and 33. For example at the section $S_1$ during early summer cruise, the depths of $\sigma = 24, 27, 30$ and 33 levels lie between 30 to 120 m, 250 to 350 m, 650 to 700 m and around 1250 m, respectively (Fig. 2).

To compare the computation results for different physical parameters we carried out a number of experiments. The symbol CA-i-j denotes a particular experiment. $i = 1, 2$ denote the early summer and autumn cruises, respectively, in 1989. $j = 1$ indicates that the inequality constraints are not imposed and $j = 2$ the inequality constraints are imposed.

From Eqns (9)-(11), the values of $T_l$ for different reference levels are computed. For the early summer data the minimum value of $T_l$ is found at $H$, the water depth, and thus the reference level is taken to be the water depth $H$. For the autumn data the minimum of $T_l$ is at $z = 600$ m level. However, since the value of $T_l$ at $z = H$ is close to the minimum value, in order to compare the results of the two data sets, the reference level for the autumn data is also taken at the water depth.

For the two values of vertical diffusion coefficient $K_c$ (1 and 10 cm$^2$/s), the respective computed volume transports differ by less than 4%. The computed volume transport for different values of vertical eddy coefficient $A_v$ (50 and 100 cm$^2$/s) changes less. In the following discussion the results are all based on $A_v = 100$ cm$^2$/s and $K_c = 10$ cm$^2$/s.

Because of the shallowness of the surface layer and also the lack of surface boundary conditions for salt, consideration of the salt convection-diffusion equation for the surface layer or not usually has no significant effect on the computation results. Our experiments confirm it, too. It is found that the two cases where the salt equation for the surface layer is considered or not result in relative changes of volume transports by less than 0.1%. Hence all results presented in the following discussions are based on consideration of the salt convection-diffusion equations for all layers except the surface layer.

On the numerical computation of this method, we discuss the following questions.

Table 1 lists the balances of the volume, salt and heat transports by the modified inverse method and those by the dynamic method, respectively. For meaningful comparison, the level of no motion for the dynamic method is also taken to be the water depth $H$. Box-1 is bounded by sections $S_i$ and $R_i$ of the early
summer observation in 1989 (Fig. 1(a)) and box-2 is bounded by the same sections of the autumn observation in 1989 (Fig. 1(b)). Table 1 indicates that (1) for the dynamic method conservations of volume, salt and heat transport are all not satisfied; (2) conservations of both volume and salt are satisfied approximately for CA-i-1 and CA-i-2.

According to the data given by Institute of Oceanography and Geography of Academia Sinica (1977), the bounds of the magnitude of the heat flux at the sea surface for boxes 1 and 2 is $10^2$–$10^3 W$. For case CA-i-1 which does not impose the inequalities (8), the unbalanced amount of the heat flux is of the order of $10^3$–$10^4 W$ (Table 1), much larger than the bounds of the heat flux at the sea surface given above. This means that the heat flux at the sea surface can not be obtained by case CA-i-1, and the inequality constraints. Egn. (8) are necessary for obtaining approximately the heat flux at the sea surface.

### Table 1. The balance of the volume, salt and heat transports in box-1 during early summer of 1989 and box-2 during autumn of 1989.

<table>
<thead>
<tr>
<th>Box</th>
<th>Method</th>
<th>$\Delta_v$</th>
<th>$\Delta_s/34.5%$</th>
<th>$\Delta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Box-1</td>
<td>CA-i-1</td>
<td>-1.72</td>
<td>-1.69</td>
<td>-8.34</td>
</tr>
<tr>
<td></td>
<td>CA-i-2</td>
<td>-0.02</td>
<td>0.00</td>
<td>-5.91</td>
</tr>
<tr>
<td></td>
<td></td>
<td>-0.02</td>
<td>-0.01</td>
<td>-0.72</td>
</tr>
<tr>
<td>Box-2</td>
<td>CA-2-1</td>
<td>0.04</td>
<td>-3.85</td>
<td>-29.51</td>
</tr>
<tr>
<td></td>
<td>CA-2-2</td>
<td>0.04</td>
<td>-0.03</td>
<td>-12.85</td>
</tr>
</tbody>
</table>

Note:

*: The level of no motion is set at the water depth $H$.

$\Delta_v$: balance of volume transport into the box (positive into the box, units: $10^3 m^3/s$).

$\Delta_s$: balance of salt transport into the box (positive into the box, units of $\Delta_s/34.5\%$: $10^3 m^3/s$).

$\Delta$: balance of heat transport into the box (positive into the box, units: $10^3 W$).

In the following we compare the results obtained with the modified method (CA-i-1 and CA-i-2) and the dynamic method.

Table 2 shows that the difference in VT obtained with the modified method (CA-i-1 and CA-i-2) and dynamic method is not large. However, the differences in VT at the lateral sections west of either box-1 or box-2 is large. For example, the VT through the lateral section west of box-1 is $6.0 \times 10^3 m^3/s$ for case CA-i-2 (Fig. 5), but $2.0 \times 10^3 m^3/s$ with the dynamic method. It is worthy of note that the Ekman transport at the section is maximum when the wind direction is parallel to this section. During early summer of 1989, the angle between the orientation of section $S_i$ ($R_f$ and $R_i$) and the wind direction is large. Thus the wind stress has little effect on the VT of the Kuroshio. However it has important effects on the VT at the lateral sections of the boxes, especially the lateral section on the shallow shelf. Table 3 shows that there are some differences in VT obtained with the modified method (CA-i-1 and CA-i-2) and dynamic method for the autumn observation of 1989.

To see the effect of the imposed surface heat flux inequality we compare the results of CA-i-1 and CA-i-2 (Tables 2 and 3). The relative changes of volume transport (VT) and heat transport (HT) between CA-i-1 and CA-i-2 are both less than 2% for the data of early summer of 1989 (Table 2). For the data of autumn of
Table 2. Comparison of the VT (10^5 m^3/s) and HT(10^8 W) at sections obtained with the dynamic method and the modified inverse method (CA-1-1 and CA-1-2) during early summer of 1989.

<table>
<thead>
<tr>
<th>Method</th>
<th>section S1 VT</th>
<th>section S1 HT</th>
<th>section R1 VT</th>
<th>section R1 HT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dynamic</td>
<td>29.2</td>
<td>28.8</td>
<td>36.1</td>
<td></td>
</tr>
<tr>
<td>CA-1-1</td>
<td>30.3</td>
<td>2.28</td>
<td>29.1</td>
<td>2.09</td>
</tr>
<tr>
<td>CA-1-2</td>
<td>30.4</td>
<td>2.31</td>
<td>28.7</td>
<td>2.06</td>
</tr>
</tbody>
</table>

*: The level of on motion is set at the water depth H.

Table 3. Comparison of the VT(10^5 m^3/s) and HT(10^8 W) at sections obtained with the dynamic method and the modified inverse method (CA-2-1 and CA-2-2) during autumn of 1989.

<table>
<thead>
<tr>
<th>Method</th>
<th>section S1 VT</th>
<th>section S1 HT</th>
<th>section R1 VT</th>
<th>section R1 HT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dynamic</td>
<td>11.4</td>
<td>33.0</td>
<td>21.2</td>
<td>30.6</td>
</tr>
<tr>
<td>CA-2-1</td>
<td>15.3</td>
<td>1.43</td>
<td>22.1</td>
<td>18.3</td>
</tr>
<tr>
<td>CA-2-2</td>
<td>16.0</td>
<td>1.51</td>
<td>29.2</td>
<td>17.6</td>
</tr>
</tbody>
</table>

*: The level of on motion is set at the water depth H.

Table 4. The horizontal and vertical volume transports through the section or the face in each layer of box-1 for early summer in 1989

<table>
<thead>
<tr>
<th>layer</th>
<th>section S1</th>
<th>(EL)_1</th>
<th>section R1</th>
<th>(WL)_1</th>
<th>(VVT)_1</th>
<th>△_j</th>
</tr>
</thead>
<tbody>
<tr>
<td>1st</td>
<td>6.94</td>
<td>-0.73</td>
<td>-6.73</td>
<td>0.25</td>
<td>0</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2nd</td>
<td>15.40</td>
<td>-2.81</td>
<td>-11.86</td>
<td>0.39</td>
<td>-0.24</td>
<td>-0.03</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>-0.87</td>
<td></td>
</tr>
<tr>
<td>3rd</td>
<td>7.59</td>
<td>-0.75</td>
<td>-8.55</td>
<td>0</td>
<td>0.87</td>
<td>-0.01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.85</td>
<td></td>
</tr>
<tr>
<td>4th</td>
<td>0.42</td>
<td>1.28</td>
<td>1.30</td>
<td>0</td>
<td>-0.85</td>
<td>-0.00</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.45</td>
<td></td>
</tr>
<tr>
<td>5th</td>
<td>0.05</td>
<td>0.64</td>
<td>0.25</td>
<td>0</td>
<td>-0.45</td>
<td>-0.01</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>total</td>
<td>30.40</td>
<td>2.37</td>
<td>28.69</td>
<td>0.64</td>
<td>0</td>
<td>-0.02</td>
</tr>
</tbody>
</table>

Note
(EL): the western lateral section of box-1
(WL): the eastern lateral section of box-1
(VVT): the upper and lower numbers denote the vertical volume transports at the upper and lower faces of the j-th layer, respectively.
△_j: the balance of volume transport into the j-th of box-1.
(positive into the layer, units: 10^5 m^3/s).

1989, however, although relative changes of both VT and HT between CA-2-1 and CA-2-2 are small for section R1, large relative change of HT reaches 9% at both sections R1 and S1 (Table 3). At section S1, the relative change of VT between CA-2-1 and CA-2-2 is as high as 24%. This indicates that the data of autumn of 1989 is probably noisier than the data of early summer of 1989. In the following, our discussions will concentrate mainly on results with inequality constraints, (8) imposed.

Now we discuss the horizontal and vertical
Table 5. The horizontal and vertical volume transports through the section or the face in each layer of box-2 for the early summer in 1989

<table>
<thead>
<tr>
<th>layer</th>
<th>section PN</th>
<th>(EL)</th>
<th>section S</th>
<th>(WL)</th>
<th>(VVT)</th>
<th>$\Delta_i$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1st</td>
<td>-7.68</td>
<td>1.22</td>
<td>6.73</td>
<td>0.12</td>
<td>0</td>
<td>-0.43</td>
</tr>
<tr>
<td>2nd</td>
<td>-17.34</td>
<td>2.84</td>
<td>11.86</td>
<td>0.42</td>
<td>0.43</td>
<td>1.81</td>
</tr>
<tr>
<td>3rd</td>
<td>-9.01</td>
<td>0.74</td>
<td>8.55</td>
<td>0</td>
<td>-1.81</td>
<td>1.53</td>
</tr>
<tr>
<td>4th</td>
<td>-0.65</td>
<td>0.63</td>
<td>1.30</td>
<td>0</td>
<td>-1.53</td>
<td>0.25</td>
</tr>
<tr>
<td>5th</td>
<td>0</td>
<td>0</td>
<td>0.25</td>
<td>0</td>
<td>-0.25</td>
<td>0</td>
</tr>
<tr>
<td>total</td>
<td>-34.68</td>
<td>5.43</td>
<td>28.69</td>
<td>0.54</td>
<td>0</td>
<td>-0.02</td>
</tr>
</tbody>
</table>

Note:
- (WL): the western lateral section of box-2
- (EL): the eastern lateral section of box-2
- (VVT): the upper and lower numbers denote the vertical volume transports at the upper and lower faces of the j-th layer, respectively.
- $\Delta_i$: The balance of volume transport into the j-th of box-2. (positive into the layer, units: $10^8$ m$^3$/s).

5. Velocity and volume transport distribution during early summer (May-June) of 1989

5.1 Velocity distribution

Fig. 3 shows the velocity distribution at section $S_i$ for case CA-1-2. The core of the Kuroshio is located over the slope. The maximum velocity is found at the 75 m level with a magnitude of 111 cm/s. As found in previous studies (e.g. Yuan and Su, 1988), in the deep layer, the axis of the Kuroshio seems to be further eastward as it flows northward. There are two countercurrents in the Kuroshio region. One is at the bottom layer near the shelf break and its maximum speed is greater than 30 cm/s. The other is below 800 m level over the lower part of the continental slope, but its speed is smaller than 2 cm/s.

At section $R_i$, the Kuroshio core is located also over the shelf break. The 100 cm/s isochron reaches as deep as 125 m and the maximum velocity is found at 50 m with a magnitude of 129 cm/s. There is no countercurrent below the Kuroshio. However, there is a countercurrent over the shelf at the bottom layer near the shelf break.

The Kuroshio speed at section $R_i$ is greater than at other sections, probably due to both the shallow water depth there and the merging of currents from the eastern lateral section of box-2 (Fig. 5). Its maximum velocity (about 149
cm/s) is found at the surface (Fig. 4). At the 200 and 400 m levels the maximum velocities are 90 and 39 cm/s, respectively, and their positions are both located east of the surface maximum. Landward of the front there is a southward current on the shelf and its maximum velocity is about 20 cm/s. Under the Kuroshio there is a very weak southward current at the bottom of the trough.

5.2 Volume transport distribution

There is a cyclonic gyre or a southward current west of Kuroshio (Fig. 4). The Kuroshio width is more than 100 km. The volume transport (VT) of the Kuroshio at sections S and R is about 30 and $35 \times 10^6$ m$^3$/s, respectively, greater than the computed Kuroshio VT during early summer of 1988 (Yuan, Su and Pan, 1991). The intrusion of the Kuroshio water over the shelf at section R is further to the west than that at sections S and R (Fig. 5). This is in agreement with the position of Kuroshio axis (Figs. 3 and 4).

Finally, as previous studies (Su and Pan, 1987; Yuan et al, 1987) pointed out, there is the northeastward current over the East China Sea shelf, which was called Taiwan Warm Current, and it may be considered as being a combination of two current systems, an inshore branch and an offshore branch. The VT of Taiwan Warm Current (TWC) which also flows through the computational region is about $1.5 \times 10^6$ m$^3$/s, i.e. $0.6 + 0.6 + 0.3 = 1.5 \times 10^6$ m$^3$/s (Fig. 5). When this shelf current flows northeastward through section R, most of it has a tendency to converge to the shelf break, similar to the findings of previous studies (e.g. Yuan and Su, 1988).

6. Velocity and volume transport distributions during autumn (October) of 1989

6.1 Velocity distribution

The velocities at section S during the 1989 autumn cruise are not very large (Fig. 6). There
are two cores at section $S_5$. One is located at the shelf break. Its maximum velocity is about 86.4 cm/s at 25 m level. The other is located over the lower part of the slope and the velocities from 25 to 125 m levels are all greater than 80 cm/s. Below the latter core there is a countercurrent and its maximum speed is greater than 10 cm/s. At the shallow end of shelf there is a southward current in the surface layer, probably due to the strong northeast winds, but near the bottom the current is still northward (Fig. 6).

Fig. 7 shows that the core of Kuroshio at section $S_5$ is located close to the shelf break and the Kuroshio front is evident. Its maximum velocity is greater than 150 cm/s at the 25 m level. There is a countercurrent occupying most part of the Okinawa Trough below 600 m. This current is rather strong; for example, its velocities at the 1000 and 1500 m levels are greater than 9 and 4 cm/s, respectively. This countercurrent extends back to section $R_1$ (Fig. 8). Its strength at section $R_1$ is even strong. Its maximum speed is greater than 22 cm/s at 800 m level, and at 1000 m level its velocity is still greater than 15 cm/s. Furthermore, the core of this countercurrent is closer to the continental slope than it is at section $S_5$. Consistent with this, the core of the Kuroshio at section $R_1$ is now placed over the shelf break (Fig. 8). The maximum velocity of the Kuroshio is about 90 cm/s at the 25 m level.

Like the early summer case, the Kuroshio Current speed at section $R_1$ is also greater than at other sections (Fig. 9), probably because of the shallow water depth at section $R_1$ and the merging of currents from the eastern lateral
Fig. 9. The velocity distribution at section $R_i$ during autumn of 1989 for case CA-2-2. (units: cm/s)

section of box-3 (Fig. 10). Its maximum velocity is about 133 cm/s at 25 m level at point 8 of $R_i$. There is a countercurrent below 950 m level, but its velocity is smaller than at other sections.

6.2 Volume transport distribution

Fig. 10 shows the distribution of volume transport in the computational region during autumn of 1989 for case CA-2-2. The volume transport through section $R_i$ is about $30 \times 10^3$ m$^3$/s. Its VT is larger than during the same seasons of 1987 and 1988. However, it is less than during early summer of 1989 (Fig. 5). The countercurrent during autumn of 1989 is rather strong. Its VT is around $3 \times 10^3$ m$^3$/s at section $R_i$ and $4 \times 10^3$ m$^3$/s at section $S_i$. However, the volume transport of the countercurrent at sections $R_i$ and $S_i$ for CA-2-1 are 4 and $7.5 \times 10^3$ m$^3$/s, respectively. Thus, the difference of the computed countercurrent VT between CA-2-1 and CA-2-2 is large for section $S_i$. We shall comment on this matter later in this paper.

Fig. 10 also shows that there is a cyclonic gyre on the shelf north of Taiwan. The temperature distribution indicates that it is a cold gyre. Fig. 10 also shows that the total VT of TWC which flows into this computational region is about $1.5 \times 10^4$ cm$^3$/s.

7. Heat transport

Table 6 lists the computed VT and HT of the Kuroshio through section $R_i$ for the early summer cruise and sections $R_i$ and $S_i$ for the autumn cruise. Other sections covered only parts of the Kuroshio and their results are not listed. For comparison the VT and HT of the Gulf Stream (Hall and Bryden, 1982) are also listed in Table 6. The symbols $Q_{1,1}$ and $Q_{1,3}$ denote the barotropic and baroclinic components of heat transport and the symbol $\bar{T}$ is defined as the average temperature:

$$\bar{T} = Q_{1,1} / (\rho C_p Q_{1,3})^{-1}$$

(18)

In general, the barotropic component of the heat transport seems to be larger than the baroclinic component for either the Kuroshio or the Gulf Stream, except for the Kuroshio at section $S_i$ during autumn of 1989 when the baroclinic component is higher than the barotropic component.

The computed results for case CA-i-2 show that during early summer of 1989 the heat transfer is from the atmosphere to the ocean for boxes 1 and 2 and their values are 7.2 and $5.7 \times 10^3$ W, respectively, totaling $1.29 \times 10^4$ W. During autumn of 1989 the total heat transfer is from the ocean to the atmosphere and it is about $2.80 \times 10^4$ W. The values of these heat exchanges at the sea surface are less than 0.5% of the HT of the Kuroshio through each section. This means that high quality hydrographic data is necessary to obtain reliable heat flux values at the sea surface.
To see the effect of the inequality constraints (8) on the computed heat flux at the sea surface, we compare case CA-i-2 with case CA-i-1 again. The computed results of case CA-i-1 show that during early summer of 1989 there is $5.65 \times 10^9$ W of heat transferred into the ocean from the atmosphere over the computational region, about 8 times larger than the computed value for case CA-i-2. The difference of the computed air-sea heat fluxes between CA-2-1 and CA-2-2 during autumn of 1989 is even larger. For example, the heat transfer from the ocean to the atmosphere through the sea surface of box-1 is $1.03 \times 10^9$ W for case CA-2-2, but $3.02 \times 10^9$ W for case CA-2-1 and in opposite direction from the atmosphere to the ocean. This seems to indicate that the quality of the 1989 autumn cruise data is not so good as the 1989 early summer cruise data and the inequality constraints, (8) is necessary for the inverse computation.

8. Summary

A modified inverse method is used to compute the Kuroshio in the East China Sea, based on the hydrographic data and moored current meter records obtained during both early summer and autumn of 1989. It is found that:

1) In general, on a broad shallow shelf the surface Ekman velocity component due to the wind stress cannot be neglected; thus our modified inverse method is useful in shallow seas. When the quality of the hydrographic data is poor, to obtain the heat flux at the sea surface, the inequality constraints at the sea surface, (8) seems to be needed.

2) The vertical volume transport (VVT) has less effect on the total volume transport of the Kuroshio. However, the VVT can not be neglected in the deep layer. The VVT is of the same order of magnitude as the horizontal volume transport through the west lateral section of box, i.e. the section on the shallow shelf in the East China Sea.

3) The volume transpots at section $R_1$ (PN) are $3.5$ and $3.1 \times 10^9$ m/s during early summer and autumn, respectively, in 1989. The heat transports at $R_1$ (PN) section are $2.5$ and $2.29 \times 10^9$ W during early summer and autumn, respectively, in 1989.

4) The Kuroshio Current speed at section $R_i$ is greater than at other sections, probably due to both the shallow water depth at section $R_i$, and the merging of currents from the eastern lateral section of box-2 (Fig. 5) or box-3 (Fig. 10).

5) A countercurrent is present in the deep layer during both cruises of 1989. It is stronger during autumn of 1989.

6) The VT of TWC flowing into this computational region is about $1.5 \times 10^9$ m/s during both cruises of 1989.

7) During both cruises of 1989 the barotropic components of HT is in general larger than the corresponding baroclinic components. About $1.29 \times 10^9$ W of heat is transferred into the ocean from the atmosphere over the computational region during early summer of 1989 and about $2.80 \times 10^9$ W of heat from the ocean to the atmosphere over the computational region during autumn of 1989.

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