Estimation and Evaluation of Sea Surface Transport, South of Japan with Sea Surface Temperature Data — Fundamental Study for Applying Satellite Data —

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Abstract

A simple sea surface transport model with an input of simulated satellite data, namely sea surface temperature, has been developed for estimating surface transport from the surface to the depth of 120 meters, south of Japan during the winter monsoon period of January to February, 1991. Given the first transport estimates from the sea surface temperature data based on the relationship between the geopotential anomaly gradient and the sea surface temperature gradient, the second order partial differential equation was solved for stream function as a non-divergent geostrophic flow under the conditions of non-slip along the coastal boundary. The estimated transport is verified by the transport from the surface to the depth of 120 meters obtained from the data of a shipboard acoustic doppler current profiler (ADCP) and the geostrophic transport estimated from the dynamic calculation along the observation lines. Results indicate that the rms error of transport between the estimated and the observed is about 2.2 Sverdrups ($10^6\text{m}^3/\text{sec}$) for the observed maximum transport of about 19 Sverdrups across the Kuroshio in the upper 120 meter layer and the streamlines derived from the transport model was qualitatively in good agreement with the flow field including the path of the Kuroshio. Since the NOAA AVHRR images received during the period of observation were not available for estimating the sea surface temperature due to the winter monsoon cloud cover, a simulated sea surface temperature image was created for analysis by spatially interpolating in situ sea surface temperature data obtained by ships.

I. Introduction

Numerous observations have been made to estimate and evaluate sea surface transport in the large boundary currents such as the Kuroshio and the Gulf Stream due to their important roles in the ocean general circulation. These observation studies are classified into two types, that

is to say “contact” and “non-contact” approach according to their different sensors and platforms used. “Contact” approach is further classified as direct (for example conventional GEK (Geomagnetic Electro Kinematograph) and an acoustic doppler current profiler (ADCP) and indirect approach (for example CTD (Conductivity, Temperature and Depth meter) for the dynamic calculation). Yamashiro *et al.* (1990) estimated the Kuroshio’s mean velocity and the maximum transport of 25 Sverdrups ($10^6\text{m}^3/\text{sec}$) in the East China Sea based on

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the 14 years' hydrographic and GEK data. Nakano *et al.* (1990) also obtained the Kuroshio transport of 24 to 28 Sverdrups across the Tokara Strait based on the data of acoustic doppler current profiler and CTD. While Richardson (1985) estimated the surface transport of 45 to 54 Sverdrups in the upper 700 m layer of the Gulf Stream using current meters and surface drifters, his estimate of the Gulf Stream transport is about twice as large as that of the Kuroshio. Taft (1972) made an geostrophic transport estimate of 50 Sverdrups in the upper 1,000 m layer in the region to the south of Japan which shows a comparable magnitude to the one estimated by Richardson. "Non-contact" approach is to estimate the surface transport based on remote sensing such as satellite altimetry and scatterometry. In 1982 Munk and Wunsch contemplated in their paper (1982) that systems of satellite altimetry and scatterometry will be sufficiently developed by the 1990s for solving the global observation problems. They stated that although the system of satellite altimetry and scatterometry give only the surface information it is possible to obtain enough information to estimate the volume transport of the large boundary currents through application of a dynamic equation. For example, Kelly and Gille (1990) made use of Geosat altimeter data in estimating the Gulf Stream surface transport and obtained the comparable magnitude with the one estimated by Richardson (1985). Although Kelly and Gille (1990) successfully demonstrated the removal of the influence of the earth geoid, the effect of uncertainties of the earth geoid on the accuracy of geostrophic velocity field cannot be tolerated. Furthermore, in the region outside the boundary currents where there is a weak geostrophic balance and is amenable to wind influences it is afraid that the altimetric

Fig. 1 Study area and observation lines with overlaid grid (0.5° longitude $\times$ 0.5° latitude) during the period of R/V Hakuhoumaru KH-91-1 research cruise (Jan. 11-Feb. 5, 1991)

measurements give a smaller estimate of geostrophic velocity than the observed one. Thus it is desirable to develop a method to estimate a sea surface transport of a wide area including boundary currents with the use of operational satellite data.

The objective of this study is to simulate satellite observations of sea surface transport, south of Japan and to validate the result based on in situ observations with a shipboard acoustic doppler current profiler.

II. In situ data and transport balance

1) In situ and simulated satellite data

The oceanographic data for this study are obtained during the period of KH-91-1 R/V Hakuhoumaru research cruise from January 11th to February 5th, 1991 covering the study area to the south of Japanese mainland (Fig. 1). Over 20 NOAA AVHRR HRPT images were

Fig. 2 The observed versus spatially interpolated sea surface temperature with different weighting functions along the observation line
Refer to the text for the description of R.

taken with an onboard receiving facility during the cruise, however none of the AVHRR images are available for analysis due to the winter monsoon cloud cover. Therefore simulated sea surface temperature (sst) images are created by spatially interpolating in situ data as follows. Based on the sst data from January 11th to February 5th collected by R/V Hakuhoumaru and the Hydrographic Department of Maritime Safety Agency, the following spatial interpolating equation (Thiebaut and Pedder, 1987, pp. 79-127) is applied.

$$X^a = \sum W(S_j) \cdot X_j / \sum W(S_j) \quad (1)$$

where X^a is an interpolated variable and X_j is an observed variable at j th location. $W(S_j)$ is a weighting distance function at distance S_j from the nearest gridpoint and is given as follows.

$$W(S) = (1 - R)/R \quad (2)$$

$$W(S) = \text{Log}(1/R) \quad (3)$$

$$R = S/d \quad (4)$$

where d is the influence radius which is a distance from the center of a circle in which the observed points are considered for interpolation

and S is the distance between the gridpoint and the observed location. Fig. 2 shows the comparison of interpolated sst with the observed sst along the observation line of R/V Hakuhoumaru. It is found that the Log $(1/R)$ function causes the interpolated sst to be over-smoothed while the $((1-R)/R)$ function causes the interpolated sst to approximate fairly well with the maximum and minimum of the observed sst. Consequently we selected the latter for interpolating sst with 0.5 degree longitude by 0.5 degree latitude. Plate 1 illustrates the interpolated sst field which clearly indicates the path of the Kuroshio with high sst and the location of cold water mass with low sst off Toukai District.

2) Sea surface transport of the Kuroshio and its balance

An acoustic doppler current profiler (ADCP) onboard R/V Hakuhoumaru makes it possible to sample simultaneously current velocities at three different depths every minute. During the cruise these measurement depths are set for 20, 75 and 120 meters so that the transport from

Plate 1 Spatially interpolated sea surface temperature field during the period of R/V Hakuhoumaru KH-91-1 research cruise (Jan. 11-Feb. 5, 1991)

Fig. 3 Distribution of horizontal transport normal to the observation lines in the upper 120 m layer

the surface to 120 m depth can be calculated. In order to evaluate the transport balance and the measurement errors of ADCP we adopt a closed study area surrounded by the observation lines and a coastal boundary in which the transport across the boundary is expected to be balanced. Fig. 3 illustrates the distribution of surface transport normal to the observation lines with characteristic points A through N.

Table 1 Volume transport in the upper 120 m layer across the observational section as to the location of A-B, B-C etc. Refer to Fig. 3

Positive transport means the inflow to the study area, while negative transport means the outflow from the study area.

Section	Transport	Subsection	Transport
KO-OM	+ 10.5	A-B	- 0.3
		B-C	+ 19.2
		C-D	- 8.4
OM-OG	- 4.5	D-E	- 5.4
		E-F	+ 4.2
		F-G	- 3.3
OG-TO	- 9.9	G-H	+ 1.2
		H-I	- 3.65
		I-J	+ 5.7
		J-K	- 8.7
		K-L	+ 7.35
		L-M	- 12.3
TOTAL	- 3.9	M-N	+ 0.5

(unit : Sverdrup)

Table 1 lists the integrated transport for each section corresponding to the characteristic points in Fig. 3. According to Table 1, the transport balance of the study area is -3.9 Sverdrups ($10^6 \text{ m}^3/\text{sec}$) which corresponds to the velocity of about 0.01 m/sec . Since the velocity measurement accuracy of the ADCP is about 0.1 Knot (0.05 m/sec), the magnitude of the transport balance is considered to be negligible and the present transports along the observation lines are adopted for further analysis and verification. In Fig. 3 the inflow of the Kuroshio can be easily seen between points B and C along the OM-KO observation line and the magnitude of the transport is 19.2 Sverdrups in the upper 120 meter layer. Taft (1972) estimated the Kuroshio's geostrophic transport in the upper $1,000 \text{ m}$ layer to be about 50 Sverdrups, thus the magnitude of transport in the upper 120 m layer corresponds to 38.4% of the Taft's estimate. Higashi *et al.* (1989) also obtained the Kuroshio's surface transport of 13.6 Sverdrups in the upper 101 m layer utilizing the shipborne ADCP which is the comparable magnitude with our estimate. Along the OG-TO observation line the section JK and LM are thought to be the Kuroshio's outflow from the study area. The sum of these two section's transport is 21 Sverdrups which is about the same magnitude as the inflow of 19.2 Sverdrups.

III. Transport model

1) Derivation of the model

Kouzai and Tsuchiya (1990) have demonstrated the possibility of applying the relationship between geopotential anomaly gradients and sea surface temperature gradients for estimating the surface current velocity of the Kuroshio based on a single NOAA AVHRR image. According to their discussion, a geostrophic

surface current velocity u off Shikoku during the period of winter (January to March) based on 10 year's data is described as

$$u = \Delta\phi / f * L \quad (5)$$

$$\Delta\phi = a * \Delta T + b, \quad a = 1.71 \text{ m}^2 \text{s}^{-2} \text{C}^{-1},$$

$$b = 0.38 \text{ m}^2 \text{s}^{-2} \quad (6)$$

where $\Delta\phi$ is the geopotential anomaly difference, f is Coriolis parameter as a constant for the study area and ΔT is the temperature difference between points with a separation L . The relationship above suggests not only the relationship between the sst field and the geostrophic velocity field but also the one between the sst field and the vertically integrated quantities computed from hydrographic data, such as mass surface transport under the assumption that the sea surface layer is well mixed. Figs. 4 and 5 show the current velocity distribution with respect to three depths of ADCP and the distribution of water temperature measured with XBT (expendable bathy-thermograph) along the OM-KO observation line. From the point of OM (X 01) to the point of X 10 (32.3°N , 134°E) it was found that the upper 120 meter layer is well mixed and the corresponding velocity differences between at the 20 meters and at the 120 meters range from 0.02 to 0.09 m/sec . Although the velocity difference at the path of the Kuroshio ($32^\circ - 32.5^\circ \text{N}$) reaches 0.2 m/sec , it is noted that this difference is within the rms error of the geopotential anomaly difference $\Delta\phi$ of $0.96 \text{ m}^2 \text{s}^{-2}$ which corresponds to the geostrophic surface velocity of 0.24 m/sec according to Kouzai and Tsuchiya (1990). Thus the geostrophic surface current velocity can be representative of the vertically averaged velocity under the condition that the surface layer is well mixed. The components of mass surface transport M_x , M_y can be defined as follows.

$$M_x = \bar{\rho} * \bar{u} * D, \quad D = 120 \text{ m} \quad (7)$$

$$M_y = \bar{\rho} * \bar{v} * D, \quad D = 120 \text{ m} \quad (8)$$

Fig. 4 Current velocity distribution with respect to three depths of ADCP along the OM-KO observation line

Fig. 5 Distribution of water temperature measured with XBT along the OM-KO observation line
The horizontal scale is arbitrary.

where $\bar{\rho}$ is the mean density and $\bar{u}$, $\bar{v}$ are the vertically averaged velocities with east-west and north-south components respectively. 120 means the depth of 120 meters which is the deepest measurement depth of ADCP. When we have these vertically integrated mass transport per unit width of flow as the first

transport estimate, it was found that the large differences between the estimated and the observed transport from ADCP data exist (refer to Fig. 12(a)). Therefore we made an attempt to increase the accuracy and reliability of model by introducing the vorticity of transport and the stream function. Under an assumption of

Fig. 6 Flow of data in the transport model

non-divergent horizontal flow, the equation of continuity holds and the stream function is defined as follows.

$$\frac{\partial M_x}{\partial X} + \frac{\partial M_y}{\partial Y} = 0 \quad (9)$$

$$M_x = \frac{\partial \Psi}{\partial Y}, \quad M_y = -\frac{\partial \Psi}{\partial X} \quad (10)$$

The first transport estimates M_x and M_y are then used to solve for a stream function from

$$\frac{\partial^2 \Psi}{\partial X^2} + \frac{\partial^2 \Psi}{\partial Y^2} = \frac{\partial M_x}{\partial Y} - \frac{\partial M_y}{\partial X} = -\text{Curl} M \quad (11)$$

with the boundary condition that the flows through the coastal boundary are zero. Curl

M in the equation (11) indicates the vorticity or curl of the transport vector M in the XY -plane. Finally the non-divergent geostrophic transports are obtained from the stream function as follows.

$$\hat{M}_x = \frac{\partial \Psi}{\partial Y}, \quad \hat{M}_y = -\frac{\partial \Psi}{\partial X} \quad (12)$$

where $\hat{M}_x$ and $\hat{M}_y$ are the non-divergent geostrophic mass transports as the second estimates. Fig. 6 illustrates the flow of data in the transport model.

2) Numerical method to solve the Poisson equation

In general there are two ways of solving Equation (11) the so called Poisson equation. One is an analytical method and the other is a numerical method. We chose the latter as the SOR (Successive Over-Relaxation) method (JASME, 1988, pp. 56-63) in the finite difference method due to easiness of getting stable solutions. The finite difference equation of the Poisson equation with the second order accuracy for the grid scheme shown in Fig. 7 is given as follows.

$$\frac{\psi_{i-1,j} - 2\psi_{i,j} + \psi_{i+1,j}}{\Delta X^2} + \frac{\psi_{i,j-1} - 2\psi_{i,j} + \psi_{i,j+1}}{\Delta Y^2} = W_{i,j} \quad (13)$$

where $\psi_{i,j}$ is a stream function located in $X=i$ and $Y=j$. ΔX and ΔY are the distances between grid points in X and Y coordinates respectively. $W_{i,j}$ represents the curl at each grid point calculated from the first transport estimate. In the SOR method the calculation is repeated between the south-west corner of the study area and the north-east corner. Acceleration parameter is taken to be 1.5 as standard value and the convergence condition is set as the one ten thousandth of the original residual $R_{i,j}$. The $(n+1)$ th approximation $\psi_{i,j}$ at the grid point i, j is given as follows.

Fig. 7 Grid scheme used for SOR method
 $\Delta X = 0.5^\circ$ longitude, $\Delta Y = 0.5^\circ$ latitude

$$\psi_{i,j}^{(n+1)} = \psi_{i,j}^{(n)} + \frac{1.5 \times R_{i,j}}{\frac{2}{\Delta X^2} + \frac{2}{\Delta Y^2}} \quad (14)$$

IV. Results and discussion

Figures 8 and 9 illustrate how the residual of approximate solution of the Poisson equation and a stream function at a grid point off Kii Peninsula are converged. Since the condition of convergence is that the residual is less than one ten thousandth of the original residual, these figures show that it takes 108 times of successive calculation for convergence.

Based on the stream function derived from

Fig. 8 Convergence of the log of residual with the times of calculation

Fig. 9 Convergence of estimated stream function vs times of calculation at a grid point off Kii Peninsula

Fig. 10 Distribution of stream function values in the unit of 10^6 kg/sec

the Poisson equation, Fig. 10 is illustrated for the distribution of stream function as streamlines which are qualitatively in good agreement with the flow field including the path of the Kuroshio. The maximum and minimum values of stream function are seen off Kii Peninsula and off Toukai District which may correspond to the path of the Kuroshio and the cold water mass respectively. According to the definition of curl, a positive curl means that the flow is counter-clockwise and vice versa. Thus the direction of the flow should be counter-clockwise around the maximum stream function value and clockwise around the minimum stream function value.

Now it is possible to verify the estimated transport with the observed transport along the observation lines. Since the CTD measurements are also made along the observation lines in addition to the ADCP measurements, two types of observed transport exists, namely the transport derived from the ADCP and the geostrophic transport obtained from the dynamic calculation. The ADCP transport is given by vertically integrating velocities in the upper 120 meter layer and may include the errors caused by the changes of ship's direction and speed and the influence of tidal current in the vicinity of coast (Ishii *et al.*, 1986). Many studies using ADCP data have been conducted to evaluate the tidal component and thereby to remove the tidal effect from ADCP data (Geyer and Signell, 1990; Foreman and Freeland, 1991). In this study we remove the errors caused by the changes of ship's direction and speed however the influence of tidal current is neglected. Instead of evaluating the tidal effect, we adopt the criterion that the transport balance of the closed study area is negligibly small. On the other hand, the geostrophic transport is obtained by vertically integrating

geostrophic velocities in the upper 120 meter layer which are given by the dynamic calculation under an assumption of a level of no-motion at 1,000 meter depth. In comparison with the ADCP transport, the geostrophic transport has the advantages over the ADCP transport in that the tidal effect is negligibly small and that the average transport is given along the observation lines. Fig. 11 is a scatterplot matrix for the two types of observed and the estimated transports along the whole observation lines. The characteristic points are summarized as follows.

(1) The rms errors between the observed and the estimated transports are 2.2 Sverdrups ($10^6 \text{ m}^3/\text{sec}$) for ADCP and 2.1 Sverdrups for geostrophic transport, which shows little differences between the two errors.

(2) There exists a good correspondence between the ADCP and the geostrophic transport with regard to all observation lines.

Fig. 12 shows the comparison of the observed with the estimated transports along each observation line. In Fig. 12 (a) the first transport estimation derived from density, geostrophic velocity and depth are given in addition to the second transport estimation by the model. In the vicinity of the path of the Kuroshio (from 31° to 32.5°N), the second transport estimation becomes closer to the observed one than the first transport estimation, which clearly indicates the validity of the transport model. Since a geostrophic equation holds for non-divergent flow, the model can not be applied for the region of meandering flow where divergence or convergence exists. For such a region another model including an effect of divergence will be desirable.

In this study an assumption is made that the sea surface temperature gradient as an input to the transport model is the same as the one

Fig. 11 Scatterplot matrix for the observed (ADCP), the calculated (dynamic calculation) and the estimated transport along the observation lines

derived from the satellite sea surface temperature (refer to Fig. 4). Since the sea surface temperature sensed by a satellite radiometer represents the one with only a few micrometers depth of the surface, it is cautioned that the applicability of satellite-derived sea surface temperature gradient for this model should be examined beforehand. Kouzai (1991) reported that the effects of sea surface micro layer on the interpretation of satellite-derived sea surface temperature are extremely complex with regard to the various parameters involved and the variability in the space-time scale.

V. Concluding remarks

The foregoing analysis leads to the following conclusion. The proposed transport model based on a simulated sea surface temperature data is a promising approach for an estimation of the sea surface transport in the region south of Japan. In conclusion the major points of this

study are summarized as follows.

(1) The model for estimating sea surface transport with an input of sea surface temperature field is developed and verified in reference to the transport obtained from the ADCP and the dynamic calculation.

(2) The rms errors between the observed and the estimated are 2.2 Sverdrups ($10^6 \text{ m}^3/\text{sec}$) for the ADCP transport and 2.1 Sverdrups for the geostrophic transport respectively. These values are considered to be reasonably small in comparison with the observed maximum of 19.2 Sverdrups in the Kuroshio, which in turn indicates the appropriateness of assuming the non-divergent geostrophic transport of the model.

(3) The rms error between the ADCP transport and the geostrophic transport is 1.5 Sverdrups along the observation lines. The good correspondence between these two transports can be seen with regard to all observation lines.

Fig. 12(a) Comparison of the observed transport with the estimated transport along the OM-KO observation line

Fig. 12(b) Comparison of the observed transport with the estimated transport along the OM-OG observation line

Fig. 12(c) Comparison of the observed transport with the estimated transport along the OG-TO observation line

(4) The transport balance of the closed study area based on the ADCP transport is -3.9 Sverdrups which is considered to be small enough in comparison with the velocity measurement accuracy of ADCP.

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海面水温データによる日本南方海域表層流量の推定と評価

——衛星データの利用をめざす基礎研究——

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海面水温場より海面から深さ 120 メートルまでの表層流量を推定するモデルを開発し、1991年1月から2月にかけて本州南方海域における船舶観測データから得た流量にもとづき検証を行なった。モデルはまず海面水温傾度と海面勾配との関係から1次推定流量を求めた後、非発散水平流れの仮定のもとでポアソン方程式を数値的に解くことによりモデルとしての2次推定流量を求めた。得られた流量を ADCP (Acoustic Doppler Current Profiler (音響ドップラー流速プロファイラー)) および力学計算により得られた海面から深さ 120 メートルまでの地衡流量にもとづき検証した。ADCP による流量および地衡流量との二乗平

均平方根 (rms) 誤差はそれぞれ 2.2, 2.1 スベルドラップ ($10^6 \text{m}^3/\text{sec}$) となり、対象海域への海面から深さ 120 メートルまでの黒潮流入量約 19 スベルドラップに比較して充分小さな値である。この方法により海面水温場から深さ 120 メートルまでの表層流量の推定が可能であることがわかった。またモデルより得られた流れ関数の分布は、黒潮流軸を含む本州南方海域の流れの場と定性的によく一致した。なお船舶による観測期間中に受信した衛星画像は冬期季節風による雲のために海面水温分布推定には利用できなかったため、観測船から観測した海面水温から内、外挿するための式を導き、この式により擬似衛星海面水温分布図を作成して利用した。

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