1. Introduction

Based on limited data sets, studies show that eddies in the South China Sea (SCS) are predominantly cyclonic in winter and anti-cyclonic in summer, with sizes ranging small to meso scale (Huang et al., 1994). Both cold and warm eddies exist in SCS. Dale (1956) and Uda and Nakao (1974) reported a cold eddy off the central Vietnamese Coast in summer. Nitan (1970) found a cold eddy located at the northwest of Luzon. Reports from the South China Sea Institute of Oceanology (SCSIO, 1985) indicate that in the central SCS, a warm-core eddy appears in summer and winter, but more close to Vietnam in summer at the surface. Recently, a cold-core eddy was detected in the central SCS during 29 December 1993 to 5 January 1994 from the analysis of TOPEX/ Poseidon data (Soong, et al., 1995).

Based on the U.S. Navy's Master Oceanographic Observation Data Set (MOOIDS), a SCS warm-core eddy with sea surface temperature (SST) higher than 29.5°C, recently reported by Chu and Chang (1995a,b), appears in the central South China Sea (west to the Lozón Island) in boreal spring, and strengthens until the onset of the summer monsoon (mid-May), then weakens and disappears at the end of May. Although its size and intensity varies, the warm-core eddy releases large moisture and heat fluxes into the atmosphere and in turn affects the monsoon circulation. Chu et al. (1995, this issue) investigated the eddy transient features and found that a warm-core eddy often appears in boreal spring during the non-El Nino years, and a cold-core eddy usually shows up during the El Nino years.

2. Combined Wind-Topographic Effect for the Central SCS Warm-Core Eddy Formation in Boreal Spring

The South China Sea (SCS) has a bottom topography (Fig.1) that makes it a unique semi-enclosed ocean basin that is overlaid by a pronounced monsoon surface wind. Extended continental shelves (less than 100 m deep) are found on the western and southern parts, while steep slopes with almost no shelves are found in the eastern part of SCS. The deepest water is confined to a bowl-type trench. The maximum depth is around 5,000 m.

Such a bowl-type bottom topography provides a precondition for the eddy formation. If the surface wind stress curl over the central SCS is anticyclonic, the Ekman downwelling prevails the central part of the 'bowl' and the Ekman upwelling appears at the boundary of the 'bowl' due to the mass balance. The downwelling prevents the deep cold water advected upward, and the upwelling helps the deep cold water advected upward. This process causes the formation of a warm-core eddy. If the surface wind stress curl over the central SCS is cyclonic, on the other hand, the combined wind-topographic effect causes the formation of a cold-core eddy.

From late winter to spring, an anticyclone appears in the central SCS from the ensemble mean atmospheric surface streamline analysis. This anticyclone generates downwelling in the central SCS and in turn prevent the cold deep water being advected to the surface. This will promote the formation of a warm-core eddy in the central SCS. As the central SCS warm-core eddy persists, being around 1°C warmer than the surroundings, an atmospheric surface low pressure center will be generated above the warm water, which may trigger the summer monsoon onset. After the summer monsoon onset, a atmospheric surface cyclone will occupy the central SCS. This cyclone will generate upwelling in the central SCS, which entrains the deep cold water into the surface mixed layer. This upwelling effect will finally destroy the central SCS warm-core eddy (Fig.2).

During El Nino events, the Asian monsoon circulation is greatly weakened (Philander, 1990), and so as the downwelling effect. Therefore, the warm-core eddy does not likely to occur in boreal winter and spring during El Nino events.

3 Primitive Equation Model

3.1 Model Description

We use a numerical model to investigate the formation of the central SCS warm-core eddy. Through the simulation, we hope to verify the wind-topographic effect. The model we use is the three dimensional model developed by Blumberg and Meller (1983, 1987) with the hydrostatic and Boussinesq approximations (Bryan, 1969) and has the following features: (1) horizontal curvilinear coordinates and an "Arakawa C" scheme (Arakawa and Lamb, 1977), (2) sigma coordinates in the vertical with realistic bathymetry, (3) a free surface, (4) a second-order turbulence closure model for the vertical viscosity (Meller and Yamada, 1974, ...
(5) horizontal diffusivity coefficients calculated by the Samagorinsky (1963) parameterization, and (6) split time steps for barotropic and baroclinic modes.

The model domain is $99^\circ E-121^\circ E$, $3^\circ S-5^\circ N$ which includes the South China Sea and the Gulf of Thailand. With its 20 km horizontal resolution and 23 vertical sigma coordinate levels, it has $125 \times 162 \times 23$ grid points (Fig. 3). All depths less than 10 m in the domain are set to be 10 m in order to satisfy the constraint $H + \eta > 0$. In vertical, the domain is divided into 23 levels $\sigma_n$ ($n = 1, 2, ..., 23$). Here $\sigma_n = n\Delta \sigma, \Delta \sigma = -1/23$. In the simulation, we use the Samagorinsky parameterization to compute the horizontal mixing coefficients with $C=0.1$, which results in the horizontal viscosity ranging from 200 to 500 m$^2$s$^{-1}$. The background vertical mixing coefficient is $10^2$ s$^{-1}$ (Blumberg and Mellor, 1983, 1987).

3.2 Lateral Boundary Conditions

There are several major straits connecting the South China Sea to the Pacific Ocean and surrounding seas. Quite a few straits are very shallow (i.e., the Strait of Malacca) or irregular (i.e., the Balabac Strait) and hard to be handled in a numerical model. Therefore, we close the Strait of Malacca, and all the small straits between Luzon Island and Borneo in the numerical model. For combine the Karimata Strait and the Gasper Strait into one open boundary. After this treatment, our model has three open boundaries: northern boundary (Taiwan Strait), eastern boundary (Luzon Strait), and southern boundary (Karimata-Gasper Strait). The transports at the three open boundaries were determined according to Wyrski’s (1961) estimation with some modification due to the boundary treatment. Table 1 shows the seasonal variation of the transport ($1\text{ sv} \equiv 10^6\text{m}^3\text{s}^{-1}$) at the three open boundaries. The summation of the total transport through the open boundaries is always zero in order to keep the mass conservation.

Table 1 The bi-monthly variation of mass transport (sv) at the open boundaries. The values were taken from Wyrski (1961).

<table>
<thead>
<tr>
<th>Month</th>
<th>Feb</th>
<th>Apr</th>
<th>Jun</th>
<th>Aug</th>
<th>Oct</th>
<th>Dec</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gasper &amp; Karimata Strait (Eastward Positive)</td>
<td>4.4</td>
<td>0.0</td>
<td>0.0</td>
<td>-0.6</td>
<td>1.0</td>
<td>4.3</td>
</tr>
<tr>
<td>Luzon Strait (Eastward Positive)</td>
<td>-1.5</td>
<td>0.0</td>
<td>1.0</td>
<td>2.5</td>
<td>-0.6</td>
<td>0.4</td>
</tr>
<tr>
<td>Formosa Strait (Northward Positive)</td>
<td>-4.9</td>
<td>0.0</td>
<td>0.0</td>
<td>0.5</td>
<td>-0.4</td>
<td>-0.9</td>
</tr>
</tbody>
</table>

The barotropic velocity at the open boundaries ($V_n^{BT}$) are obtained by

$$V_n^{BT} = \frac{\text{Transport}}{\int l dh}$$

(1)

where $l$ is the boundary width, and $n$ means the normal direction to the boundary. The baroclinic velocity, $V_n^{BC}$, is calculated by
\[ V^{BC}_n(\sigma) = V^{BT}_n \frac{\epsilon}{1 - \epsilon^{-1}} \]  
\[ \frac{\partial}{\partial t}(T, S) + V_n \frac{\partial}{\partial n}(T, S) = 0 \]

During the numerical integration, both \( V^{BT}_n \) and \( V^{BC}_n \) were interpolated into the time step from the bi-monthly values of the volume transport listed in Table 1.

On the lateral open boundaries, temperature (T) and salinity (S) are prescribed at inflow boundaries, whereas at outflow boundaries, the advection equation, namely,
\[ \frac{\partial}{\partial t}(T, S) + V_n \frac{\partial}{\partial n}(T, S) = 0 \]
is solved for both barotropic and baroclinic modes.

3.3 Stability Constraints

The stability constraints were derived from various linearized subsets of equations (Blumberg and Mellor, 1987). Since \( \Delta x = 20 \text{ km, } H_{\text{max}} = 5000 \text{ m, and } U_{\text{max}} = 1 \text{ m/s, the typical CFL condition derived from Blumberg and Mellor (1987) for the external mode is } 45 \text{ s and for the internal mode is } 2000 \text{ s. In our numerical model, the time step was set as } 25 \text{ s for the external mode and as } 900 \text{ s for the internal mode.}

3.4 Surface Forcing

The atmospheric forcing includes wind forcing and thermodynamic forcing. We took the monthly mean climatological wind stress data (Hellerman and Rosenstein, 1983) as the wind data for the middle of the month and interpolated them on each day. The wind stress has a typical magnitude of 1 dyne/cm² (corresponding to surface wind speed 10 m/s). Northeastwinds dominate during the winter monsoon season, and southwestwinds prevail during the summer monsoon season. Such a wind stress field is called the climatological wind forcing.

There are two approaches setting up surface thermal forcing: (1) flux forcing when the surface heat and salt fluxes, \( Q_H (W \text{ m}^{-2}) \) and \( Q_S (m \text{ s}^{-1}) \) are given, and (2) restoring forcing when the fluxes are not given and the model surface temperature (T) and salinity (S) are relaxed to the observed values \( (T_{obs}) \) and \( (S_{obs}) \):
\[ K \frac{\partial T}{\partial z} = \alpha_1 Q_H + \alpha_2 C(T_{obs} - T), \]
\[ K \frac{\partial S}{\partial z} = \alpha_1 Q_S + \alpha_2 C(S_{obs} - S), \]
where \( K \) is the thermal exchange coefficient and \( C \) is the relaxation constant taken to be 0.7 m/d, which is equivalent to a relaxation time of 43 days for an upper layer of 30 m thick. \( (\alpha_1, \alpha_2) \) are (0,1)-type switcher parameters: \( \alpha_1 = 1, \alpha_2 = 0 \), we only have the flux forcing; \( \alpha_1 = 0, \alpha_2 = 1 \), we only have the restoring forcing; \( \alpha_1 = 1, \alpha_2 = 1 \), we have both the flux forcing and the restoring forcing. In this simulation we only use the restoring forcing as the surface thermal forcing. We construct the \( (T_{obs}, S_{obs}) \)
data through interpolation from the monthly mean temperature and salinity data (Levitus, 1982). Both horizontal and vertical resolutions of the Levitus (T,S) data are different from our model. We use the optimal interpolation (OI) to convert the Levitus (T,S) data on our model grids.

4. Model Simulation Results

The model was initiated from April mean (T,S) fields (Levitus, 1982), and forced by the surface wind stress (Hellerman and Rosenstein, 1983) and thermodynamical fluxes depicted in (6). Both the surface forcing and the transport at the open boundaries were interpolated into daily data. The model were integrated for three years. The third year’s outputs were used for discussion.

Three days of the simulated surface currents were picked up for investigating time evolution of the circulation pattern. We choose 30 January for the winter monsoon forcing, 30 March for the boreal spring forcing, and 30 April for the pre-summer monsoon forcing.

On 30 January, there is a strong southward coastal jet appearing near the Vietnam coast, and there is no cyclone or anticyclone in the central SCS (Fig.4a). The isolines of SST in the central SCS almost parallel to the southeast China coast (Fig.5a). Neither warm-core eddy nor cold-core eddy is found on that day.

On 30 March, the strength of the southward Vietnam coastal currents is greatly reduced. An anticyclone with a strong northward branch is generated in 13°N-18°N, 111°E-115°E (Fig.4b). Besides, several small cyclones are also simulated, e.g., a small and weak one (next to the anticyclone) is located at 13°E-16°E, 115°E-116°E. A warm-core eddy is also simulated associate with the anticyclone (Fig.5b).

On 30 April, the southward coastal jet disappears from near the Vietnam coast. The SCS an anticyclone persists in 13°N-18°N, 111°E-115°E (Fig.4c). A warm-core eddy is also simulated associate with the anticyclone (Fig.5c).

The simulations confirm our scenario for the warm-core eddy formation under the climatological forcing. The cold-core eddy formation during El Nino events should be studied in the future research.

5. Conclusions

A hypothesis for the SCS eddy formation under the climatological forcing (more likely non-El Nino years) was also proposed in this study. During late winter an atmospheric anticyclone over the central SCS generates downwelling and in turn prevent the cold deep water being advected to the surface. As solar radiation increases in the late winter monsoon, the central SCS surface water is heated more than the surrounding waters by the downwelling effect. This will promote the formation of a warm-core eddy in the central SCS. As the central SCS warm-core eddy...
Figure 4: Model simulated surface circulations for (a) 30 January, (b) 30 March, and (c) 30 April.

Figure 5: Model simulated SST for (a) 30 January, (b) 30 March, and (c) 30 April.
persists with few degree warmer than the surroundings, an atmospheric surface low pressure center will be generated above the warm water, which may trigger the summer monsoon onset. After the summer monsoon onset, a atmospheric surface cyclone occupying the central SCS. This cyclone will generate upwelling in the central SCS, which sucks the deep cold water into the surface mixed layer. This upwelling effect will finally destroy the central SCS warm-core eddy. This downwelling hypothesis on the SCS warm-core eddy formation during non El Nino years were verified by a 3-D primitive equation model. The air-sea feedback scenario has not been verified for El Nino events.

We recommend that further study should be (1) studying 3-D thermal features of the SCS mesoscale eddies; (2) constructing a composite for the SCS surface wind fields, to see if it is an upwelling or downwelling favorable wind field; (3) running the 3-D numerical model under the El Nino wind forcing.

Acknowledgments

Authors are grateful to D.S. Ko for invaluable comments. This work was funded by the Naval Oceanographic Office, the Office of Naval Research NOMP Program, and the Naval Postgraduate School.

References