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NOTES AND CORRESPONDENCE

Modeling and Verification of the Subsurface Current Core of the Ryukyu Current

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ABSTRACT

In recent years, several observations based on direct current measurements and model simulations have been successful in detecting a northeastward undercurrent, "the Ryukyu Current", along the Pacific side of the Ryukyu Islands with a unique "subsurface northeastward current core" structure. The volume transport (20 - 25 Sv) of the Ryukyu current completes the volume transport budget (45 - 50 Sv) of the Kuroshio system. A Pacific Ocean circulation model based on the RIAM Ocean Model (RIAMOM) with 1/12 horizontal resolution successfully reproduced the observed structures of the northeastward Ryukyu Current with a subsurface core at 500 - 600 m. A three-layer model simulation shows the existence of the Ryukyu Current and explains the mechanism of subsurface current maximum through blocking effect of bottom topography around the Ryukyu Islands.

Key words: Ryukyu current, Subsurface current core, RIAMOM, Three layer model

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1. INTRODUCTION

Since the volume transport of the Kuroshio in the ECS is only half (20 - 25 Sv) (Nakamura and Hyuga 1999; Ichikawa and Beardsley 2002) of the Kuroshio south of Japan (40 - 45 Sv) (Imawaki et al. 2001), many researchers have been thinking for a long time that there should be a northeastward current, the Ryukyu Current (hereafter, the RC), along the Pacific side of the Ryukyu Islands (hereafter, RI) which supplies the missing transport to the Kuroshio.

In recent years, several observations based on direct measurements using Inverted Echo Sounders with pressure gauges (PIES), moored current meters or ADCPs combined with hydrographic observations have been successful in detecting a current with a unique structure of the RC along the RI (Yuan et al. 1995; Zhu et al. 2003, 2005, 2006; Ichikawa et al. 2004; Konda 2005;).

In a model study, Nakamura et al. (2007) performed numerical experiments using two primitive equation models that incorporate realistic and idealized topography, respectively. They showed that bottom intensification of the RC is formed due to the first baroclinic mode topographic Rossby wave that emanates from the Kuroshio in the Tokara Strait. As Nakamura et al. (2007) indicated, further high resolution models are needed to reproduce a realistic RC system.

You and Yoon (2004) successfully reproduced the observed structure of the RC along the Pacific side of RI with a subsurface current core at 500 - 600 m using 1/6 Pacific Ocean circulation model. You and Yoon (2004) showed that the volume transport of the RC increases flowing northeastward along the RI from about 5.6 Sv near the strait east of Taiwan to 20.5 Sv near the Tokara Strait. The volume transport near the Tokara strait was large enough to explain the missing transport of the Kuroshio. The study of You and Yoon (2004) showed that the northeastward increase of the volume transport along the RI was supplied by the westward

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Subtropical Current. You and Yoon (2004), however, did not explain the formation mechanism of the subsurface current core of the RC.

Building on the work of You and Yoon (2004), this paper describes the structure and variability of the RC in the 1/12 Pacific Ocean circulation model, compares them with observations and tries to clarify the structure of the RC all the way along the RI. A hypothesis to explain the formation mechanism of the RC with subsurface current core is verified using a three-layer model, focusing on the blocking effect of the RI.

2. MODEL

2.1 Ocean General Circulation Model

The RIAM Ocean Model (RIAMOM) used in this study is a primitive general ocean circulation model with a free surface developed by Lee and Yoon (1994) at the Research Institute for Applied Mechanics (RIAM). Detailed explanations of the model were described in You and Yoon (2004) and You (2005).

The model covers the Pacific Ocean from 95 E to 70 W and from 50 S to 65 N (Fig. 1). The horizontal grid intervals are 1/12 in both latitudinal and longitudinal directions and the number of vertical levels is 70. The model bottom topography is based on the National Geophysical Data Center ETOPO5 with 5-min resolution. Biharmonic horizontal diffusion is used for both momentum and tracers. The coefficient is 8.0 10^{17} cm^{4} s^{-1} for momentum, and that for tracers is 8.0 10^{16} cm^{4} s^{-1} for this study.

The model is integrated from a state of rest with climatological mean temperature and salinity distribution of World Ocean Atlas (WOA) 94 (Levitus and Boyer 1994; Levitus et al. 1994), and forced by the monthly mean NCEP wind stress during the period from 1979 to 2001. To prescribe the heat flux at the surface, a combined boundary condition (Barnier et al. 1995) is used with net heat flux of NCEP and the sea surface temperature data of WOA 94 (Levitus and Boyer 1994). For the surface forcing (momentum and heat flux), climatological monthly mean data are linearly interpolated to provide the value at each time step. The surface salinity is restored to the climatological value of seasonal salinity data of WOA 94 (Levitus et al. 1994) with a restoring time scale of 10 days. The model is integrated for 25 years, and the last five years are analyzed.

2.2 Three Layer Model

The layered model used in this study is a non-linear primitive equation ocean model. The model can treat more than 2 layers with free surface. No barotropic/baroclinic mode-splitting method is adopted, so a short time step for Courant-Friedrichs-Lewy (CFL) condition is required to solve the surface gravity wave. The model physics is basically the same as those of Holland and Lin (1975) and Kim and Yoon (1996).

The model domain is assumed to be rectangular, mid-latitude ocean with 0.5 in both latitudinal and longitudinal directions and 3 vertical layers (Fig. 2). To incorporate the topographic effect, an inclined island chain similar to the RI is designed in the western part of the simple rectangular ocean as shown in Fig. 2. The first layer thickness is assumed to be about 500 m and the bottom of the second and

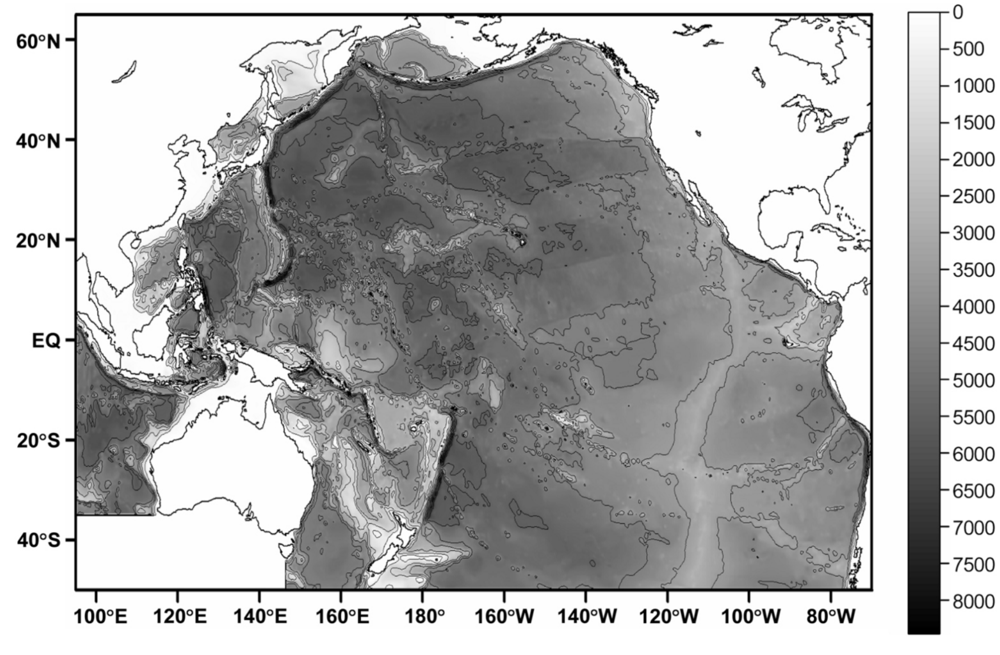


Fig. 1. Model domain and topography. Depths are in meters. The contour interval is 1000 m.

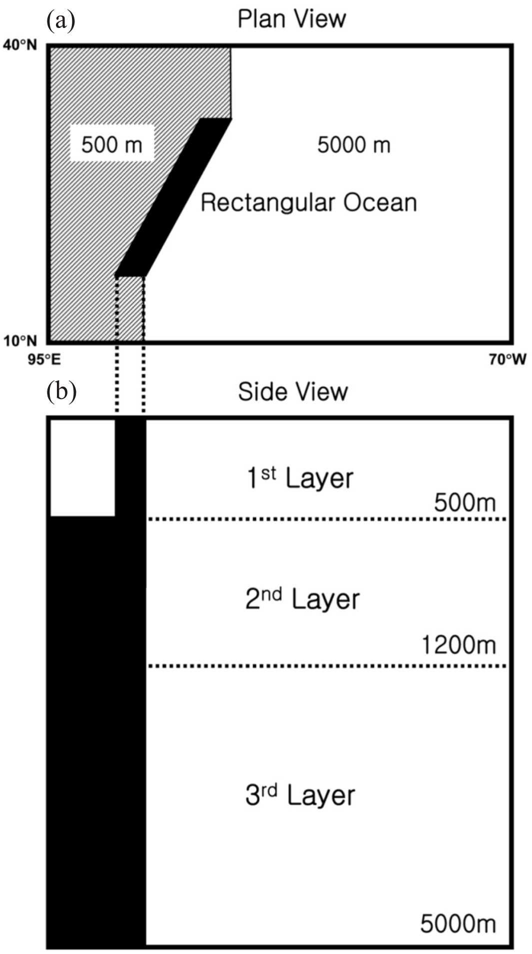


Fig. 2. Schematic views of three-layer model domain and thickness. The black marking indicates the islands and bottom topography.

third layer is assumed to be about 1200 and 5000 m, respectively. Seawater densities of each layer are 1.017, 1.024, and 1.027 g cm^{-3} which are averaged values from previous OGCM results, respectively. *R* (Rayleigh drag coefficient), A_h (Horizontal eddy viscosity coefficient), and (Newtonian damping coefficient) are 1.0 10^{-7} sec^{-1}, 1.0 10^{8} cm^{2} sec^{-1}, and 7.0 10^{-7} sec^{-1} in the present model, respectively. As for the wind forcing, we used simple wind stress curl which is maximum of 1 at 40 N and minimum of -1 at 10 N [Eq. (1)] to generate subtropical gyre. The equations are

$$\tau^{\lambda} = \sin [\pi \quad (j\ grid\ number - total\ j\ grid/2)/total\ j\ grid]$$
$$\tau^{\phi} = 0 \qquad (1)$$

where λ latitude and ϕ longitude.

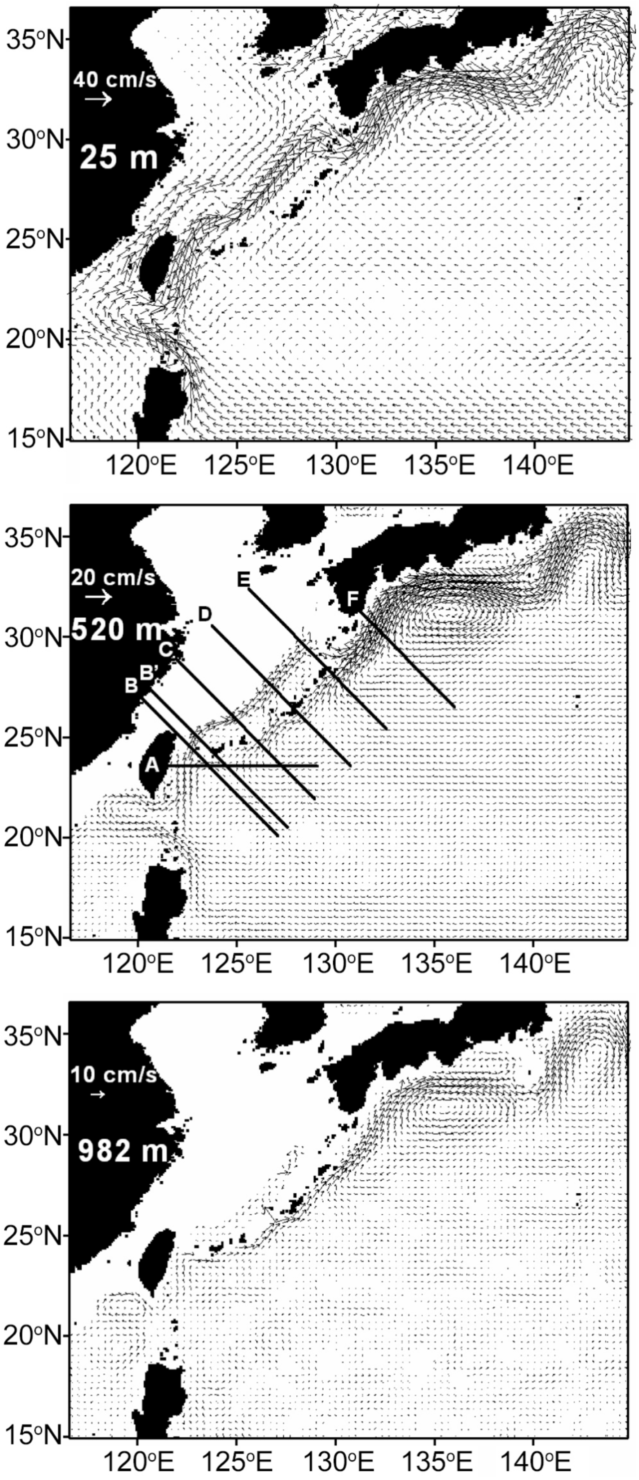


Fig. 3. Mean velocity fields at 25, 520, and 982 m depth.

3. RESULTS

3.1 Mean Fields

Figure 3 shows modeled current fields averaged for the last 5 years at 25, 520, and 982 m. At 25 m depth, the Kuroshio passes mainly through the strait east of Taiwan into the ECS and flows along the ECS continental shelf break toward the Tokara Strait. At 520 m depth, a remarkable northeastward flow, the RC, can be seen along the Pacific side of the RI as well as the Kuroshio along the ECS continental shelf break. At 982 m depth, the RC along the

Pacific side is much stronger than the Kuroshio along the ECS continental shelf break. Similar to You and Yoon (2004), in this 1/12 Pacific Ocean circulation model, we can identify another western boundary current, the RC, flows along the Pacific side of the RI.

Vertical sections of the 5-year mean horizontal current normal to lines A through F in Fig. 3 are shown in Fig. 4. Vertical structures of the current fields from this study are very similar to those of You and Yoon (2004). The Kuroshio with a typical structure of a western boundary current at the A-line enters mainly into the ECS through the strait east of Taiwan (B-line) and flows northeastward along the continental shelf break with a large volume transport. The other northeastward flow, the RC, can be seen along the Pacific side of the RI with a subsurface current core at 500 - 600 m depth (B through E lines). The subsurface current core of the RC is very weak at the B-line, and then becomes stronger as it moves northeastward along the RI (C through E lines). At the C and D-lines the maximum core velocities of the northeastward flow of the RC are 10 and 15 cm s^{-1}, respectively. And then at the E-line the maximum current core of the RC increases to about 35 cm s^{-1}. Eventually, the Kuroshio in the ECS and the RC merge at the Tokara Strait and feed the Kuroshio south of Japan (F-line). These velocity fields of Fig. 4 reveal that the RC, the northeastward western boundary current, is characterized by a subsurface velocity core. It originates from the Pacific side of the RI around east of Taiwan and becomes stronger as it flows along the RI, implying a volume transport supply from somewhere.

The stream function field of volume transport in Fig. 5 and the schematic view of the volume transport budget in Fig. 6 clarify the volume transport supply to the RC. The volume transport of the RC is about 8.0 Sv at the B -line and increases to about 10.6 Sv at the D-line (Okinawa Island) and to about 18.5 Sv at the E-line (Amami-Ohshima Island). You and Yoon (2004) showed that the volume transport of the RC is about 5.7 Sv at its origin (same as B -line) near the strait east of Taiwan and gradually increases to about 15.5 Sv at Okinawa and about 21.3 Sv at Amami-Ohshima. Therefore, this study shows that the volume transport at its origin

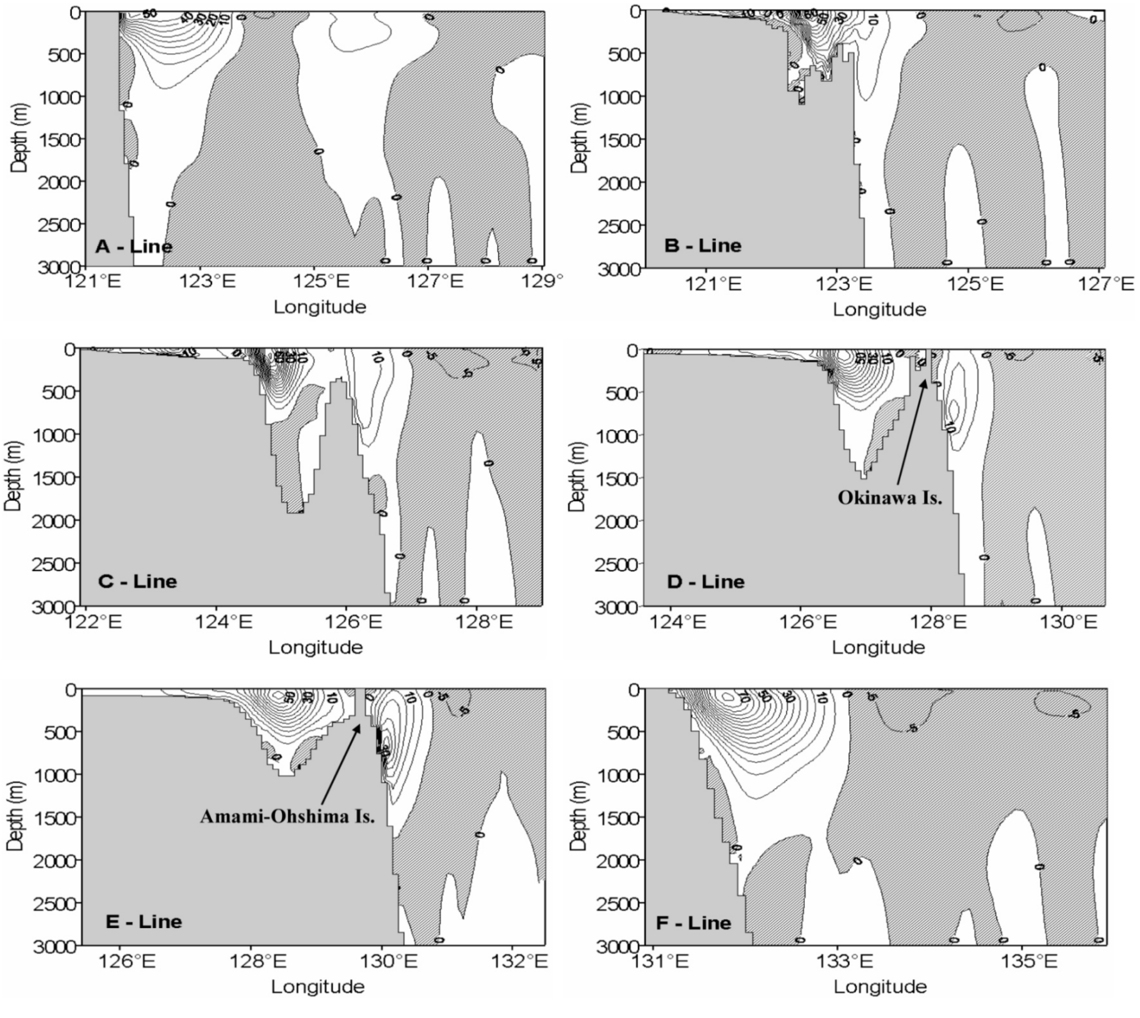


Fig. 4. Vertical sections of horizontal velocities normal to sections from A to F lines for the Kuroshio system around Ryukyu Islands. The contour interval is 5 cm s^{-1}. Shading indicates southwestward velocity (B to F line) and southward velocity (A line).

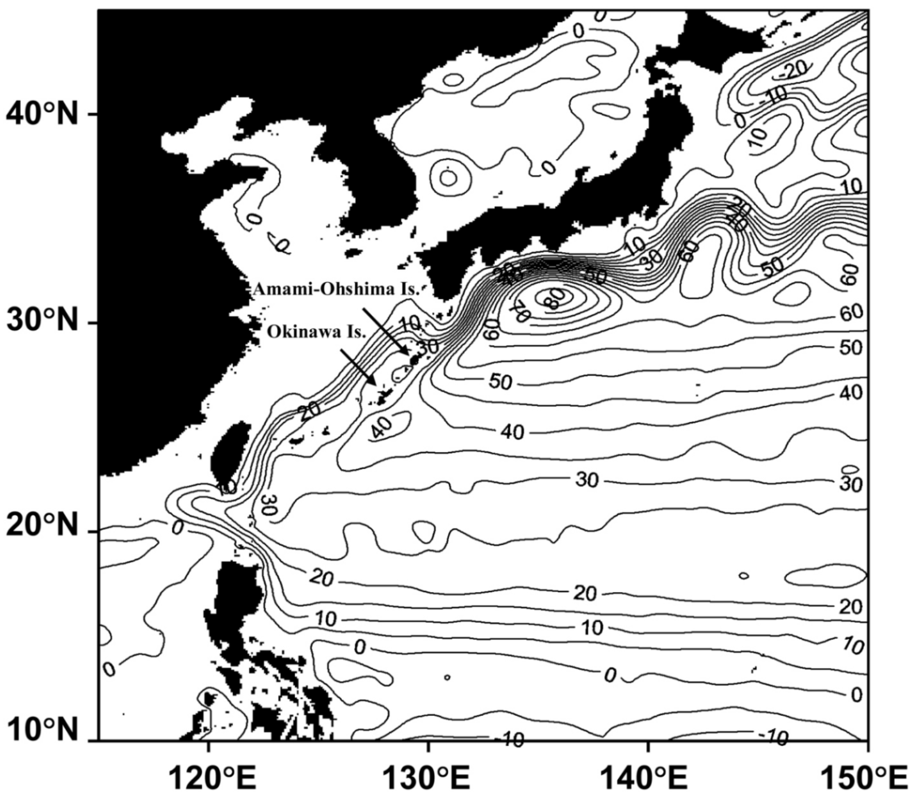


Fig. 5. Five-year mean volume transport stream function upper 1500 m depth in the northwestern Pacific Ocean. The contour interval is 5 Sv.

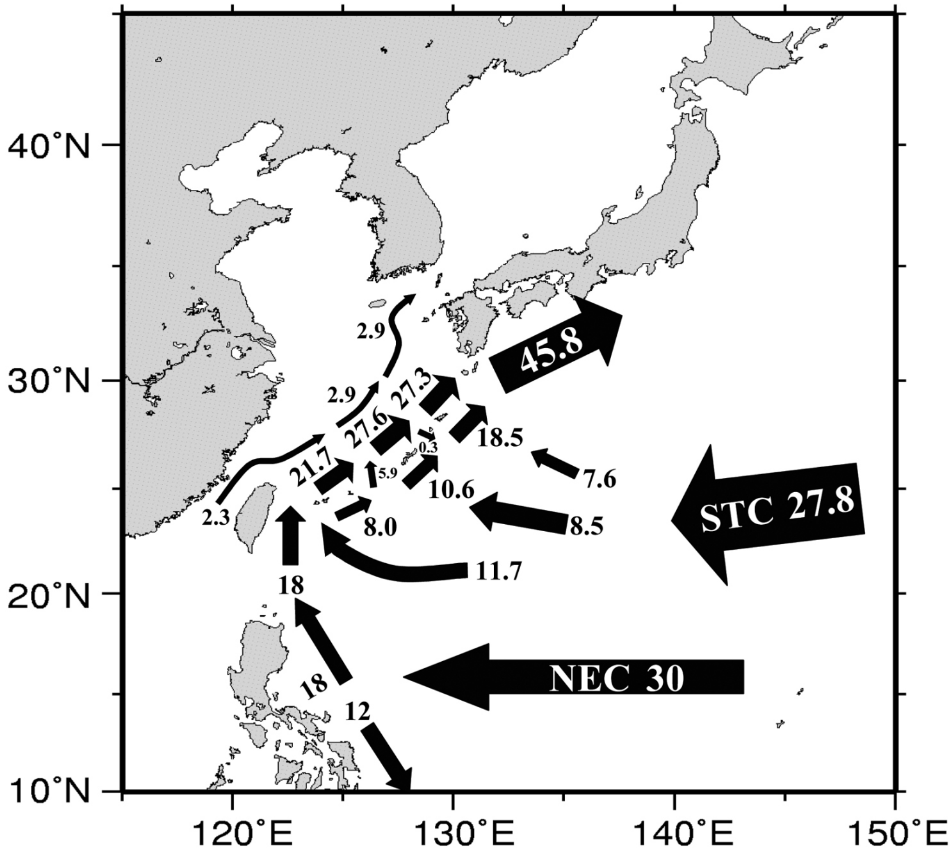


Fig. 6. Schematic representation of the volume transport in the northwestern Pacific Ocean based on the stream function in Fig. 5.

of B -line is about 2.3 Sv larger than that of You and Yoon (2004). However, at Okinawa and Amami-Ohshima Islands this study shows smaller volume transport of 4.9 and 2.8 Sv, respectively, than that of You and Yoon (2004). The increase of volume transport along the RI supports recent observations that the volume transport of the RC increases from 6.1 Sv southeast of Okinawa (Zhu et al. 2003) to 18 - 20 Sv southeast of Amami-Ohshima Island (Ichikawa et al. 2004). You and Yoon (2004) showed that the increase of volume transport is supplied by a broad westward flow with a relatively strong magnitude, the “Subtropical Current (STC)” in the upper 500 m between 20 and 26°N in the western Pacific Ocean.

The Kuroshio with a volume transport of 27.3 Sv in the ECS and the RC with a volume transport of 18.5 Sv merge at the Tokara Strait, feeding the Kuroshio south of Japan with a net volume transport of 45.8 Sv. The Kuroshio volume transport of 27.3 Sv in the ECS is in good agreement with observed values (Nakamura and Hyuga 1999; Ichikawa and Beardsley 2002).

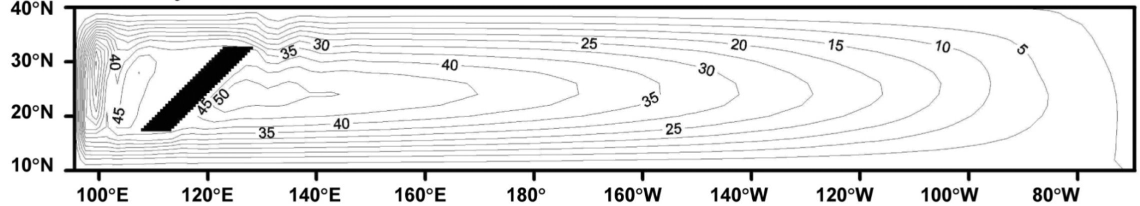
3.2 Formation of the Subsurface Current Core

We assume the existence of island chains such as the RI to be a key factor for the formation of subsurface current core. To investigate the effect of the RI, we simulated the subsurface current core using a three-layer model with simple wind forcing. The blocking effects of these islands are considered to be of primary importance for subsurface core formation. A subtropical gyre is assumed to be driven by an external force in an idealized rectangular ocean with three layers as seen in Fig. 2. The first layer thickness is assumed to be slightly larger than the depth (about 500 m) of the Tokara Strait and the strait east of Taiwan. The tops of the second and third layers in motion are assumed to be always below the depth of these straits.

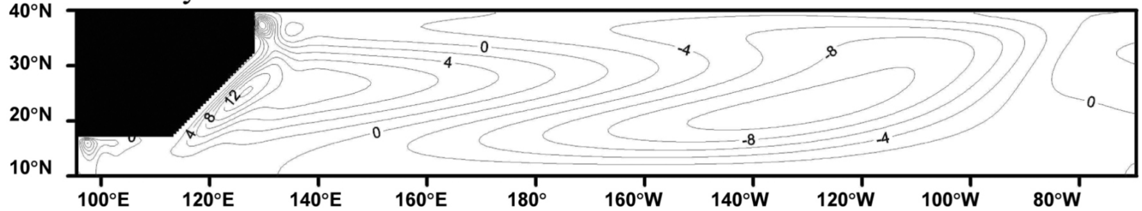
The subtropical clockwise circulation with a western intensification is generated due to the westward intensification (-effect) and nonlinear effect. In the real ocean, the RI has two main straits, the strait east of Taiwan and the Tokara Strait. These Straits may allow the upper portion of the western boundary current shallower than the strait sill (about 500 m) to shift westward due to the -effect so that the remaining western boundary current of the first layer in the Pacific side of the RI becomes weak enough compared with the western boundary current in the second layer (compare Figs. 7a and b). In the first layer (Fig. 7a), volume transport of the western boundary current east of the islands is about 5 Sv which corresponds to that of the RC.

In the second layer (Fig. 7b), clockwise circulation weaker than in the first layer is generated. Then the western boundary current, that is, RC can flow along the Pacific side of the RI with a subsurface current core showing a maximum volume transport of about 12 Sv as the Kuroshio south of Japan. In the third layer (Fig. 7c) countercurrent along the

(a) 1st layer



(b) 2nd layer



(c) 3rd layer

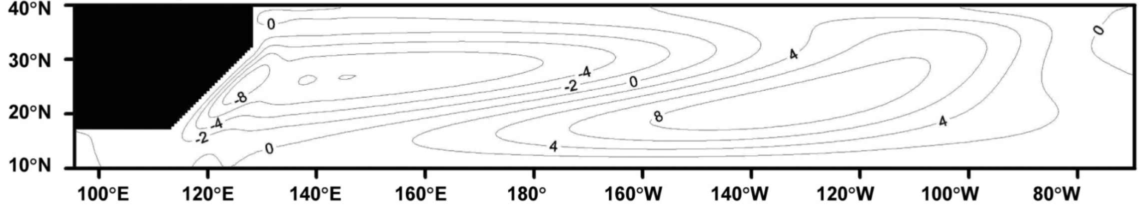


Fig. 7. Simulated streamlines for idealized subtropical ocean of (a) 1st layer, (b) 2nd layer, and (c) 3rd layer. The contour interval is 5 Sv (1st layer) and 2 Sv (2nd and 3rd layer).

RI shows a maximum volume transport of about -8 Sv so that we cannot identify the remarkable western boundary current east of the islands.

This is based on the assumption that the introduction of two straits may not change so much the western boundary current in the second layer because the second layer is deeper than the sill depth of the straits so that the western boundary current at the deeper layer is blocked by the bottom topography. In each instance, it appears that the blocking effects of a shallow ridge associated with an island chain allow only the upper portion of the flow to be transported into the marginal sea. As a result, the remaining western boundary current on the eastern side of the island chain has a relatively weak upper-layer current and a subsurface maximum.

4. CONCLUSION

Results of 1/12 Pacific Ocean Model successfully captured observed structures of the RC along the Pacific side of the RI with subsurface current core at about 500 - 600 m depth. The 5-year mean volume transport of the RC in the model increases flowing northeastward along the RI from 8.0 Sv near the strait east of Taiwan to 18.5 Sv near the Tokara Strait. The volume transport of the RC is similar to previous model results of You and Yoon (2004). The essence of these features is well reproduced by the three-layer model that successfully obtained the subsurface maximum current which increases northeastward along the RI, a result that corresponds well to the OGCM results.

This study shows the blocking effect of bottom topography around the RI as a probable mechanism to explain the formation of the subsurface current core of the RC. The three-layer model shows successfully the existence of the RC and explains the mechanism of subsurface current core. Despite recent intensive observations, even the mean structure of the RC is not fully captured due to its great fluctuations in the real ocean. Therefore, more extensive and comprehensive observations of the RI are required to detect the spatial and temporal changes of the RC.

This study presented a possible mechanism for the formation of the RC with a subsurface current maximum. The seasonal and interannual variabilities, the interaction between eddies and the RI, etc. still remain to be solved.

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