Numerical Study of the Generation and Propagation of Trigger Meanders of the Kuroshio South of Japan

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We examine the processes underlying the generation and propagation of the small meander of the Kuroshio south of Japan which occurs prior to the transition from the non-large meander path to the large meander path. The study proceeds numerically by using a two-layer, flat-bottom, quasi-geostrophic inflow-outflow model which takes account of the coastal geometries of Kyushu, Nansei Islands, part of the East China Sea, and the Izu Ridge. The model successfully reproduces the observed generation and propagation features of what is called “trigger meander” until it passes by Cape Shiono-misaki; presumably because of the absence of the bottom topography, the applicability of the present numerical model becomes questionable after the trigger meander passes by Cape Shiono-misaki. The generation of the trigger meander off the south-eastern coast of Kyushu is shown to be associated with the increase in the supply of cyclonic vorticity by the enhanced current velocity in the upper layer along the southern coast of Kyushu where the no-slip boundary condition is employed. Thereafter, the trigger meander propagates eastward while inducing an anticyclone-cyclone pair in the lower layer. The lower-layer cyclone induced in this way, in particular, plays a crucial role in intensifying the trigger meander trough via cross-stream advection in the upper layer; the intensified trigger meander trough then further amplifies the lower-layer cyclone. This joint evolution of the upper-layer meander trough and the lower-layer cyclone indicates that baroclinic instability is the dominant mechanism underlying the rapid amplification of the eastward propagating trigger meander.

1. Introduction

It is well known that the Kuroshio path south of Japan exhibits remarkable bimodal features, namely, the large meander (LM) path and the non-large meander (NLM) path (see Fig. 1). These bimodal features cannot be found in other western boundary currents such as the Gulf Stream. The Kuroshio path variations exert a strong influence on fisheries, ship navigation, and marine resources so that the bimodal features of the Kuroshio have attracted the attention of many researchers for a long time.

One of the important observed features is that the transition from the NLM path to the LM path is preceded by the amplification of a small meander during the course of the eastward propagation after generation off the south-eastern coast of Kyushu. The small meander is therefore often called the trigger meander (Solomon, 1978). By analyzing the time series of sea level difference between Naze and Nishinoomote (see Fig. 1) from 1965 to 1992, Kawabe (1995) showed that the Kuroshio current velocity through the Tokara Strait once increased before the trigger meander was formed, and, except for 1975, the trigger meander was generated while thus increased Kuroshio current velocity was relaxing. Although several numerical studies have been carried out to investigate the transition from the NLM path to the LM path preceded by the eastward propagation of the trigger meander (Sekine and Toba, 1981; Chao, 1984; Yasuda et al., 1985; Yoon and Yasuda, 1987; Akitomo et al., 1991), such time-dependent features observed in the Tokara Strait have not been taken into account. An exception is the numerical study by Akitomo et al. (1997), but their study is limited to the barotropic responses of the Kuroshio. Considering that the time variation of sea level difference between

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Naze and Nishinoomote mostly reflects that of the surface geostrophic current velocity through the Tokara Strait, the study of baroclinic responses of the Kuroshio is indispensable as well.

This motivated us to carry out numerical experiments using a two-layer quasi-geostrophic (QG) model. In the present study we first demonstrate that the simple two-layer, flat-bottom, inflow-outflow model can reproduce the observed generation and propagation features of the trigger meander remarkably well, until the trigger meander passes by Cape Shiono-misaki. Next, by analyzing the calculated results, we show that lateral shear of the Kuroshio is relevant to the generation of the trigger meander off the south-eastern coast of Kyushu, whereas the effect of vertical current shear is essential in the amplification of the eastward propagating trigger meander. It should be noted that, presumably because of the absence of the bottom topography, the applicability of the present numerical model becomes questionable after the trigger meander passes by Cape Shiono-misaki so that the present study does not discuss the stationary path of the Kuroshio including the transition from the NLM path to the LM path.

2. Model

The basic equations are the QG equations for a two-layer ocean given by

\[
\frac{\partial}{\partial t} \left( \nabla_{h}^{2} \psi_1 + \frac{f o}{D_1} h \right) \\
= - \frac{f o}{D_1} J(\psi_1, h) - J(\psi_1, f + \nabla_{h}^{2} \psi_1) + A_{H} \nabla_{h}^{4} \psi_1, \quad (1)
\]

\[
\frac{\partial}{\partial t} \left( \nabla_{h}^{2} \psi_2 - \frac{f o}{D_2} h \right) \\
= \frac{f o}{D_2} J(\psi_2, h) - J(\psi_2, f + \nabla_{h}^{2} \psi_2) + A_{H} \nabla_{h}^{4} \psi_2, \quad (2)
\]

where \( \psi_1 \) and \( \psi_2 \) are the stream functions in the upper and lower layers, respectively, defined by

\[
u_i = - \frac{\partial \psi_i}{\partial y}, \quad \psi_i = \frac{\partial \psi_i}{\partial x} \quad (i = 1, 2) \quad (3)
\]

where \((u_i, v_i)\) are the corresponding horizontal current velocities; the upper and lower layer thicknesses in the undisturbed state, \(D_1\) and \(D_2\), are assumed to be 500 m and 3500 m, respectively; the horizontal eddy viscosity coefficient \(A_H\) is assumed to be 500 m²s⁻¹; \(J(A, B) = \frac{\partial A}{\partial x} \cdot \frac{\partial B}{\partial y} - \frac{\partial A}{\partial y} \cdot \frac{\partial B}{\partial x}\) is the Jacobian operator; \(f\) is the Coriolis parameter defined by
where $f_0 = 7 \times 10^{-5}$ s$^{-1}$, $\beta = 2 \times 10^{-11}$ m$^{-1}$s$^{-1}$, and taking account of the effect of the tilted coastline from Kyushu to the Izu Ridge, we rotate the $(x, y)$ axes counter clockwise by $\theta = 20$ degrees from the eastward and the northward directions, respectively (Chao, 1984; Yoon and Yasuda, 1987; Yamagata and Umatani, 1989; Sekine, 1990; Akitomo et al., 1991, 1997); and $h$ is the interface displacement defined by

$$h = \frac{f_0}{g} (\psi_2 - \psi_1),$$

where the reduced gravity $g^*$ is 0.02 ms$^{-2}$. The QG equations (1) and (2) are replaced with a finite difference scheme by applying the centered difference leapfrog scheme. In particular, the Arakawa Jacobian is used for an expression of the advective term (Arakawa, 1966).

The model geometry is essentially the same as that presented in Akitomo et al. (1997) (see Fig. 2). This model basin is divided into square grids with a horizontal spacing of 10 km. The Kuroshio is driven in the upper layer by assuming inflow at the lower left-hand corner and outflow at the upper right-hand corner. At the outlet, the Kuroshio, with a triangular velocity profile 100 km wide, is assumed to have maximum current velocity $U(t)$, whereas at the inlet we assume the boundary conditions given by

$$\frac{\partial \psi_i}{\partial y} - \frac{\partial \psi_i}{\partial y} = 0 \quad (i = 1, 2).$$

As an initial state we assume a narrow jet (just inside the dash-dotted line in Fig. 2) with the same profile as that at the outlet all the way from the inlet down to the outlet, and then we let it go. Until transient responses die away and hence quasi-stationary flow is attained, $U(t)$ at the outlet is kept constant at $U_b$. Thereafter we change the value of $\psi_i$ at the coast of Nansei Islands as well as at the right-hand side and lower side boundaries such that pulse-like variation with magnitude $\Delta U$ is superposed on $U_b$ at the outlet (Fig. 3). Associated with this, pulse-like variation of the averaged Kuroshio current velocity occurs in the upper layer in the Tokara Strait. Although the variation of the Kuroshio current velocity in the Tokara Strait, reproduced in this way, might be too simple to simulate the observed results shown in Fig. 3, the same time-dependent boundary condition was also employed in the barotropic model of Akitomo et al. (1997), which makes it easier to see the difference between the barotropic and baroclinic responses of the Kuroshio.

All lateral boundaries are assumed to be no-slip except at the coast of the Izu Ridge. To prevent the computational noise generated at the inlet and the outlet from affecting the interior region, a high viscosity zone is assumed in the shaded region in Fig. 2, where the horizontal eddy viscosity coefficient $A_H$ is increased linearly from 500 up to 2500 m$^2$s$^{-1}$ as the right-hand side and lower side boundaries are approached. The numerical computation proceeds using a discrete time step of 3600 s. To avoid numerical instability, the Euler backward scheme (Matsuno, 1966) is applied five times successively at every 100 time steps.
Fig. 3. The left-hand panel shows the time series of the maximum current velocity of the Kuroshio assumed at the outlet (see Fig. 2) which corresponds to the time variations of sea level difference between Naze and Nishinoomote (for the geographical locations, see Fig. 1) observed in 1986 and 1989 shown in the right-hand panel (Kawabe, 1995). The arrow in the right-hand panel shows the phase at which the trigger meander was generated.

Fig. 4. Comparison of the streamline patterns in the upper layer for $U_b = 0.8 \text{ m s}^{-1}$ and $\Delta U = 0.55 \text{ m s}^{-1}$ with the observed flow patterns in 1989 reported by the Japan Maritime Safety Agency (Oceanographic Prompt Reports). Contour interval for the streamlines is $5 \times 10^3 \text{ m s}^{-1}$. Dashed contours are negative.
3. Results

We carried out a total of 63 numerical experiments by varying \( U_b \) from 0.4 ms\(^{-1} \) to 1.0 ms\(^{-1} \) at intervals of 0.1 ms\(^{-1} \) and \( \Delta U \) from 0.3 ms\(^{-1} \) to 0.7 ms\(^{-1} \) at intervals of 0.05 ms\(^{-1} \). For \( \Delta U < 0.4 \) ms\(^{-1} \), no appreciable small meander can be reproduced off the south-eastern coast of Kyushu, independent of the values of \( U_b \). For \( \Delta U \geq 0.4 \) ms\(^{-1} \), by contrast, the small meander generated off the south-eastern coast of Kyushu becomes prominent, although it decays during the subsequent eastward propagation unless \( U_b \geq 0.5 \) ms\(^{-1} \). In order to reproduce the generation and propagation features of the trigger meander, therefore, these parameters should be in the range of \( U_b \geq 0.5 \) ms\(^{-1} \) and \( \Delta U \geq 0.4 \) ms\(^{-1} \). As \( \Delta U \) becomes larger for each \( U_b \), the amplitude of the trigger meander generated off the south-eastern coast of Kyushu becomes larger, whereas, as \( U_b \) becomes larger for each \( \Delta U \), the trigger meander thus generated amplifies more rapidly during the subsequent eastward propagation.

As representative examples, Figs. 4 and 5 show the calculated results for \( U_b = 0.8 \) ms\(^{-1} \) and \( \Delta U = 0.55 \) ms\(^{-1} \) and for \( U_b = 0.6 \) ms\(^{-1} \) and \( \Delta U = 0.6 \) ms\(^{-1} \), which are found among others to simulate most closely the observed stream patterns for 1989 (Figs. 4a’–f’) and for 1986 (Figs. 5a’–f’) reported by the Japan Maritime Safety Agency (Oceanographic Prompt Reports). Although it is impossible to validate the chosen values of \( U_b \) with the observed results, \( \Delta U \approx 0.6 \) ms\(^{-1} \) is consistent with the observed result that the sea level difference between Naze and Nishinomote increased by about 0.3 m (Fig. 3), so long as it is caused by the enhancement of the upper-layer geostrophic current velocity through the Tokara Strait.

We now briefly describe the calculated results, checking by means of a comparison with the observed stream patterns. The phase for each streamline pattern is shown in Fig. 3. As \( U(t) \) increases and hence approaches the maximum, the Kuroshio gradually separates from the south-eastern coast of Kyushu (see Figs. 4a and 5a). The separation continues even while \( U(t) \) is decreasing, and by the end of the decelerating stage (\( t \approx 60 \) days), the trigger meander is generated off the south-eastern coast of Kyushu (Figs. 4b and 5b). The trigger meander generated in this way is seen to be amplified while propagating eastward with an anticyclonic eddy behind it (Figs. 4c–f and 5c–f). Although the agreement between the calculated results and the observed stream patterns for 1986

![Fig. 5. Comparison of the streamline patterns in the upper layer for \( U_b = 0.6 \) ms\(^{-1} \) and \( \Delta U = 0.6 \) ms\(^{-1} \) with the observed flow patterns in 1986 reported by the Japan Maritime Safety Agency (Oceanographic Prompt Reports). Contour interval for the streamlines is \( 5 \times 10^3 \) m\(^2\)s\(^{-1}\).](image-url)
is not as good as that for 1989, the general characteristics of the observed features are reproduced well as a whole, implying that the essential physics of the phenomenon is successfully simulated in the present simple two-layer model.

In the barotropic model of Akitomo et al. (1997), the trigger meander was generated dozens of days after the short-term variation of the Kuroshio current velocity was applied in the Tokara Strait, which is inconsistent with the observed results. Considering that the model geometry and initial and time-dependent boundary conditions in Akitomo et al. (1997) are essentially the same as those in the present study, this inconsistency can be attributed to the neglect of baroclinicity.

4. Discussion

Figure 6 shows the areas of relative vorticity exceeding $1 \times 10^{-5} \text{s}^{-1}$ superposed on the calculated streamline pattern at $t = 60$ days for the representative case shown in Fig. 4. A narrow band of cyclonic vorticity is seen to extend from the southern coast of Kyushu to the trough of the trigger meander, suggesting that the generation of the trigger meander is associated with the relative vorticity supplied at the southern coast of Kyushu where the no-slip boundary condition is employed. This is more directly confirmed by the fact that the generation of the trigger meander does not occur at all in the numerical experiment with all the lateral boundaries assumed to be free-slip (results not shown). Figure 7 shows the time variation of the maximum current velocity in the upper layer through the Tokara Strait, whereas Fig. 8 provides that of the term balances in Eq. (1) at the location $(x, y) = (350 \text{ km}, -210 \text{ km})$ marked by a star (★) in Fig. 6. After the current velocity in the upper layer along the southern coast of Kyushu attains the maximum at $t = 40$ days, the relative vorticity at this location rapidly increases, mainly through the advection of cyclonic vorticity from the Tokara Strait. The enhancement of the Kuroshio along
the southern coast of Kyushu thus increases the viscous production as well as downstream advection of cyclonic vorticity, resulting in the generation of the trigger meander off the south-eastern coast of Kyushu. This suggests that the same generation process might be reproduced using a reduced gravity model. In fact, the reduced gravity model with the same values of $U_b$ and $\Delta U$ as assumed in Fig. 4 reproduces the small meander of nearly the same amplitude at $t \approx 60$ days (Fig. 9).

In contrast, the propagation features of the small meander are quite different between the two-layer model and the reduced gravity model. Figure 10 shows the time variations of the total amplitude of the small meanders shown in Figs. 4 and 9, respectively. The trigger meander in the two-layer model rapidly amplifies during $t = 90$–120 days, quite different from the small meander in the reduced gravity model, which decays during the same time period.

To understand the underlying mechanism for the rapid amplification of the trigger meander reproduced in the two-layer model, we focus on the first term on the right-hand side of Eq. (1), namely, $f_0 J(\psi_1, h)/D_1$ which vanishes in a reduced gravity model. Figure 11(a) shows the contours of $f_0 J(\psi_1, h)/D_1$ (thick lines) superposed on
the $\psi_1$ fields (thin lines) for the case in Fig. 4. Until the rapid amplification of the meander trough almost terminates at $t = 110$ days (see Fig. 10), large negative values of $\psi_1, h/D_1$ can be found at the location of the steepening trough. From Eq. (1) we can see that the negative values of $\psi_1, h/D_1$ correspond to the cross-stream advection, causing partial relaxation of the interface displacement as well as vertical stretching of upper-layer parcels, accompanied by enhancement of cyclonic vorticity (Cushman-Roisin, 1994). It is interesting to note that, using Eq. (5), $\psi_1, h/D_1$ can alternatively be written as $\psi_1^2/J(\psi_1, \psi_2)/g^* h/D_1$, so that the strength of the cross-stream advection effect in the upper layer is greatest when the induced lower-layer currents flow at large angles to the upper-layer Kuroshio. In fact, Fig. 11(b) shows that the trigger meander amplifies rapidly during $t = 90–120$ days when an anticyclone-cyclone pair below the meander trough-crest pair evolves, but the trigger meander ceases to amplify after $t = 120$ days when the angle between the upper- and lower-layer currents becomes fairly small and, furthermore, the cyclone below the steepened meander trough decays, presumably through the lateral friction at the southern coast of Japan. The joint evolution of the upper-layer meander trough and the lower-layer cyclone is also confirmed to occur for the case shown in Fig. 5 (results not shown), indicating that baroclinic instability is the important mechanism underlying the rapid amplification of the eastward propagating trigger meander.

The importance of $\psi_1, h/D_1$ was first pointed out by Yoon and Yasuda (1987), who demonstrated the relevance of baroclinic instability to the amplification of the trigger meander, based on the idealized two-layer numerical model. Even after their study, however, the dynamics of transient responses of the Kuroshio and the stationary path of the Kuroshio following the transient processes have been discussed using the barotropic numerical model (e.g. Yamagata and Umatani, 1989; Akitomo et al., 1991, 1997). The main reason for this is that there has been no attempt to check the validity of each model result through comparison with the observed stream patterns. The numerical model of Yoon and Yasuda (1987), for example, was somewhat too simple to allow comparison with the observed stream patterns where the trigger meander was introduced in the upper layer a priori in the form of a cyclonic eddy, with the generation process being obscured. The emphasis of the present study is placed on the demonstration that the observed generation and propagation features of the trigger meander can be simulated very well using the two-layer, quasi-geostrophic model, which includes the coastal geometries of Kyushu, Nansei Islands, part of the East China Sea, and Izu Ridge, and, more specifically, the mechanism of baroclinic instability.

5. Concluding Remarks

Using a two-layer, flat-bottom, quasi-geostrophic numerical model, we have reproduced the generation and propagation features of the trigger meander of the Kuroshio which agree remarkably well with the observed ones until the trigger meander passes by Cape Shionomisaki. The generation of the trigger meander off the south-eastern coast of Kyushu is associated with the increase in the supply of cyclonic vorticity caused by the enhanced current velocity in the upper layer along the southern coast of Kyushu where the no-slip boundary condition is employed. Thereafter, the trigger meander propagates eastward while inducing an anticyclone-cyclone pair in the lower layer. The lower-layer cyclone induced in this way, in particular, plays a crucial role in intensifying the trigger meander trough via cross-stream advection in the upper layer; the intensified trigger meander trough then further amplifies the lower-layer cyclone. This joint evolution of the upper-layer meander trough and the lower-layer cyclone indicates that baroclinic instability is the important mechanism underlying the rapid amplification of the eastward propagating trigger meander, consistent with the results of Yoon and Yasuda (1987) based on the idealized two-layer numerical model.

Several important problems still remain to be investigated. First, the effect of bottom topography should be clarified. When a realistic bottom topography is taken into account, the Kuroshio tends to flow along a continental shelf slope instead of flowing into the offshore, deeper, flat-bottom region where it is susceptible to baroclinic instability (Sekine, 1992). To clarify the effect of bottom topography, we have to take a more precise account of the existence of density stratification. Ikeda (1983), for example, showed that at least a three-layer model should be used to study the instability of a current flowing along a continental shelf slope. The absence of the bottom topography in the present numerical model might also be responsible for the disagreement between the calculated stream pattern and the observed one after the trigger meander passes by Cape Shionomisaki. For example, we could not reproduce what is called the “S-shaped” path where the Kuroshio loops back west of the Izu Ridge just before the LM path is formed (Sakaïda et al., 1998). Since the reliability of the stationary path of the Kuroshio that follows this transient process is considered to be questionable, the present study has not discussed the stationary path of the Kuroshio including the transition from the NLM path to the LM path. Second, it is not clear what causes the short-term variation of the Kuroshio current velocity in the Tokara Strait, although the interaction between the Kuroshio and a mesoscale eddy approaching the Tokara Strait might be a plausible explanation. To answer these questions, we are planning a detailed nu-
Fig. 11. (a) Contours of $f_0 J(\psi_1, h)/D_1$ (thick lines) superposed on the $\psi_1$ fields (thin lines) for the case shown in Fig. 4. Contour intervals are $5 \times 10^{-12}$ s$^{-2}$ for $f_0 J(\psi_1, h)/D_1$ and $1 \times 10^4$ m$^2$s$^{-1}$ for $\psi_1$. The areas where $f_0 J(\psi_1, h)/D_1 \leq -5 \times 10^{-12}$ s$^{-2}$ are shaded.
(b) The $\psi_2$ fields (thick lines) superposed on the $\psi_1$ fields (thin lines) for the case shown in Fig. 4. Contour intervals are $2 \times 10^5$ m$^2$s$^{-1}$ for $\psi_2$ and $1 \times 10^4$ m$^2$s$^{-1}$ for $\psi_1$. Dashed contours are negative.
merical study based on a 3-D, primitive, high resolution level model that takes account of realistic topography, the results of which will be reported elsewhere.

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