Instability and Long-Term Variability of Strong Eastward Jet in an Oceanic Double-Gyre

Shinya SHIMOKAWA*, Tomonori Matsuura** and Hironori Hashimoto**

* National Research Institute for Earth Science and Disaster Prevention, Tsukuba, Ibaraki, Japan  
** University of Toyama, Toyama, Japan

Reynolds number (Re) dependency on the state of the oceanic double-gyre is investigated using a quasi-geostrophic model under constant forcing. The results are classified by energy level and length of the strong eastward jet. Both the energy levels and the lengths of the jet increase with an increase in Re until Re = 209, but begin to decrease at Re = 314. For Re = 209, intense eddy shedding occurs from the easternmost tip of the jet. For Re = 314, the jet is collapsed and eddy shedding occurs not only from the easternmost tip of the jet but also in the path of the jet. The realistic phenomena are considered to correspond to the results for Re = 314 in this study. The dominant instability in the long-term variability of the oceanic double-gyre seems to be different from the instability of the G-mode suggested by Simonnet and is related to shear instability.

1. INTRODUCTION

In the mid-latitude North Pacific, a westerly wind blows eastward around 45°N and a trade wind blows westward around 15°N. The water level rises around 30°N because of the southward Ekman transport by the westerly wind and northward Ekman transport by the trade wind. An ocean current, known as a (clockwise) sub-tropical gyre, is generated as a result of geostrophic balance to the increase in water level. In the same way, an ocean current, known as a (counter-clockwise) sub-polar gyre, is generated by a balance between westerly and circumpolar winds. Both gyres combined are referred to as the oceanic double-gyre. The oceanic double-gyre is one of the basic configurations of large-scale surface ocean circulation. Recently, the long-term variability (over 10 years) has been studied in relation to climate variability 1).

In recent studies on the long-term variability of the oceanic double-gyre, numerical models with middle complexity (e.g., a quasi-geostrophic model) have played a central role because they have the benefit of facilities for theoretical interpretation of the results and comparison them with realistic phenomena.

Simonnet and Dijkstra (2002) 2) suggested using a quasi-geostrophic model in which the long-term variability of the oceanic double-gyre is generated by oscillation of G-mode as a result of merging of P-mode related to the existence of the multiple steady state and L-mode related to the strength of the oceanic double-gyre. They suggested that tri-pole/symmetric characteristics in P-mode and dipole/anti-symmetric characteristics in L-mode are essentially important in the oscillation of G-mode (i.e., the long-term variability of the oceanic double-gyre). On the other hand, Berloff et al. (2007) 3) suggested using a different quasi-geostrophic model that realistic state of the oceanic double-gyre is more complicated than the state suggested by Simonnet and Dijkstra (2002) 2). They also suggested that the long-term variability of the oceanic double-gyre is generated by a turbulent oscillator as a result of interaction between changes in the gyre strength and north-south movement of the gyre boundary. They
concluded that Simonnet and Dijkstra (2002)'s result is valuable only in regions of low Reynolds number (Re) and the realistic state of the oceanic double-gyre is more turbulent (i.e., in regions of high Re).

Therefore, to clarify the long-term variability of the oceanic double-gyre, we investigate Re dependency on the state of the oceanic double-gyre using a quasi-geostrophic model under constant forcing and compare the results with the realistic phenomena. Another object of this study is to review Re dependency on the state of the oceanic double-gyre from a geophysical viewpoint because this has not been investigated systematically (at least, it has not been summarized) even under constant forcing.

2. NUMERICAL MODEL AND EXPERIMENTAL METHOD

The numerical model used in this study is a 1.5 layer, reduced-gravity, quasi-geostrophic numerical model with Rayleigh-type interfacial friction and a non-slip boundary condition (Fig. 1) \(^4\). The model domain is a rectangle of 3600 km × 2800 km representing the mid-latitude North Pacific. The forcing wind has only an east-west component, is constant in time, and is north-south varying in space, representing a simplified spatial distribution of wind stress on the mid-latitude North Pacific. The time interval of integration is \(7.2 \times 10^3\) s (2 h), and the horizontal grid spacing is \(2.0 \times 10^4\) m (20 km). Further details of the numerical model are shown in Shimokawa and Matsuura (2010) \(^5\).

Control parameters in the experiments involve only Re, defined as \(Re = \frac{2 \pi \tau_0 \rho_0 A_h \beta H}{\rho_0 \tau_0}\), where \(\tau_0\) is the amplitude of wind stress (= 0.1 N m\(^{-2}\)), \(\rho_0\) is the mean density of seawater (= 1000 kg m\(^{-3}\)), \(A_h\) is the coefficient of eddy viscosity, \(\beta\) is the north-south gradient of the Coriolis parameter (= 2×10\(^{-11}\) m\(^{-1}\) s\(^{-1}\)), and \(H\) is the reference thickness of the upper layer (= 1000 m). In the experiments, only \(A_h\) changes from 1200 m\(^2\) s\(^{-1}\) (Re = 26) to 100 m\(^2\) s\(^{-1}\) (Re = 314). We conducted the experiments over 250 years for Re = 26, 31, 39, 70, 95, 112, 157, 209, and 314.

![Figure 1 Model configuration used in this study.](image)

3. RESULTS

Figs. 2 and 3 show averaged flow pattern and time-series of total energy for each Re, respectively. For Re = 26 (Fig. 2a), the flow pattern shows a stable asymmetric pattern. For Re = 31 and 39 (Figs. 2b and 2c), an oscillation appears between an asymmetric pattern and an anti-symmetric pattern. For Re = 70 (Fig. 2d), the flow pattern shows a stable modon-like pattern \(^6\). For Re = 95 (Fig. 2e), the flow pattern changes from a modon-like pattern to a stable asymmetric pattern after around year 80 with a slight decrease of the total energy (Fig. 3e). For Re = 112 (Fig. 2f), the flow pattern becomes
turbulent and eddy shedding occurs from the easternmost tip of a strong eastward jet. For Re = 157 (Fig. 2g), the flow pattern shows a wavy variation of the jet, but eddy shedding does not occur. For Re = 209 (Fig. 2h), intense eddy shedding occurs from the easternmost tip of the jet (see also Fig. 4). For Re = 314 (Fig. 2i), the jet is collapsed and eddy shedding occurs not only from the easternmost tip of the jet but also in the path of jet (see also Fig. 5). The changes described here can be understood from the state of the total energy (Fig. 3).

The changes can be classified into four groups as follows: 1) Re = 26, 31, 39, 2) Re = 70, 95, 112, 3) Re = 157, 209 and 4) Re = 314. Each group has a similar average energy level (after the first rapid changes for each Re, for example, after around year 38 for Re = 26; see Fig. 3a). Moreover, each group has a similar length of the strong eastward jet corresponding to the energy level. That is, the energy level and the length of the jet change discontinuity (i.e., a quantization with respect to both the energy level and the length of the jet occurs).

Figure 2: Average flow pattern (height of 1st layer) for (a) Re = 26, (b) Re = 31, (c) Re = 39, (d) Re = 70, (e) Re = 95, (f) Re = 112, (g) Re = 157, (h) Re = 209 and (i) Re = 314.

Figure 3: Time series of total energy for (a) Re = 26, (b) Re = 31, (c) Re = 39, (d) Re = 70, (e) Re = 95, (f) Re = 112, (g) Re = 157, (h) Re = 209 and (i) Re = 314. The periods indicated by arrows in (h) and (i) are used in Figs. 4, 5.
4. DISCUSSIONS

The states for Re = 209 and 314 are turbulent, but it can be considered that they are different qualitatively because their energy levels and lengths of the jet are largely different. In particular, it is interesting that both the energy levels and the lengths of the jet increase with an increase in Re until Re = 209, but begin to decrease at Re = 314. Therefore, we investigate the results of Re = 209 and 314 in more detail.

Figs. 4 and 5 show changes of flow pattern in high and low energy periods for Re = 209 and 314. For Re = 209, the flow pattern is stable at a high energy (year 181) and unstable at a low energy (year 189). In the unstable state, eddy shedding occurs from the easternmost tip of the jet. Similarly, for Re = 314, the flow pattern is stable at a high energy (year 176) and unstable at a low energy (year 185). However, in the unstable state, the jet is collapsed and eddy shedding occurs not only from the easternmost tip of the jet but also in the path of the jet.
Figure 4 Flow patterns for $Re = 209$ in year 181 and year 189. The corresponding periods are indicated in Fig. 3h.

Figure 5 Flow patterns for $Re = 314$ in year 176 and year 185. The corresponding periods are indicated in Fig. 3i.

Fig. 6 show the time-series of available potential energy (APE) and kinetic energy (KE) including the periods indicated in Figs. 4 and 5 for $Re = 209$ and 314. For $Re = 209$, KE decreases with the decrease of APE and the time lag is only a few months. On the other hand, for $Re = 314$, APE shifts to KE over a long-term (about a decade) and then KE oscillates intensely. These differences are considered to represent differences in the instability. The oscillation of KE for $Re = 314$ corresponds to the collapse of the jet and subsequent intense eddy shedding shown in Fig. 5. Therefore, it is suggested that the oscillation is related to the long-term variability of the oceanic double-gyre.

Fig. 7 shows annual average sea surface height and the integrated path center (0 m line of sea surface height) in 1996 and 2002 based on MOVE/MRLCOM-WNP (Meteorological Research Institute Multivariate Ocean Variational Estimation Community Ocean Model Western North Pacific)
The state in 1996 is more turbulent than that in 2002. In 1996, eddy shedding occurs not only from the easternmost tip of the jet but also in the path of the jet, but in 2002, the state becomes stable. That is, the transition between a stable state and an unstable state occurs. This situation is considered to correspond to the results for Re = 314 in this study.

In 1996, meander in the strong eastward jet is large and the southern recirculation is weak. On the other hand, in 2002, meander in the jet is small and the southern recirculation is strong. However, the latitude of flow axis (i.e., gyre boundary) does not change significantly. Berloff et al. (2007) suggested that interaction between change in the gyre strength and north-south movement of the gyre boundary is important in long-term variability of the oceanic double-gyre. However, at least in this case, the long-term variability seems not to be related to north-south movement of the gyre boundary.

![Figure 6](image_url)

Figure 6 (a) Available potential energy and (b) kinetic energy for Re = 209 during years 176-195 and (c) available potential energy and (d) kinetic energy for Re = 314 during years 171-190.

![Figure 7](image_url)

Figure 7 Annual average of sea surface height and the integrated path center (0 m line of sea surface height) in 1996 and 2002 based on MOVE/MRI.COM-WNP.

Simonnet (2005) explained that the long-term variability of the oceanic double-gyre is generated by oscillation of G-mode as a result of merging of P-mode and L-mode even for large Re (Figs. 6-9.
of Simonnet, 2005). However, we consider that L-mode for the largest Re (lower left panel in Fig.9 of Simonnet, 2005) is clearly different from L-modes for the smaller Re. As stated in the introduction, the essential characteristic of L-mode is anti-symmetric about the center axis in the north-south direction. In contrast, L-mode for the largest Re does not have such anti-symmetry. In addition, L-mode for the largest Re shows eddy shedding in the path of the jet. Therefore, it is speculated that the case for 3-bump in Simonnet (2005) corresponds to the results for Re = 314 in this study, and the cases for 2-bump in Simonnet (2005) correspond to the results for Re = 209 in this study.

This difference can be visualized clearly in Figs. 4 and 5. For Re = 209, the jet is meandering, but eddy shedding in the path of the jet can’t be seen. In this case, the meandering of the jet is considered to be the result of the instability of G-mode. On the other hand, for Re = 314, the meandering of the jet overshoots and eddies pinch off in the path of the jet. In this case, the eddy shedding is considered to be the result of the instability.

In conclusion, the realistic phenomena are considered to correspond to the results for Re = 314 in this study, and the instability related to the long-term variability of the oceanic double-gyre seems to be different from the instability of G-mode. In this regard, our conclusion is same as Berloff et al. (2007)’s conclusion. However, as stated above, part of the long-term variability can’t be explained by the mechanism suggested by Berloff et al. (2007).

One of the reasons for the existence of such varied explanations is because the state is very turbulent and it is not easy to clarify the mechanism of the long-term variability. However, the main instability for Re = 314 is speculated to be shear instability from the difference of the flow pattern between Re = 209 and 314 (Figs.4 and 5). The necessary condition for shear instability is given by the Rayleigh-Kuo criterion: “the sign of $\text{RK} = \beta - \frac{\partial \bar{U}}{\partial y} y^2$ changes within the considered region”

where \(\bar{U}\) is an averaged east-west velocity. In calculation of RK, arbitrariness exists in the choice of period and region for the averaging of east-west velocity. In this study, 10 years and the western half of the model region are chosen as the period and region, respectively, because the target in this study is a long-term variability of the oceanic double-gyre and the strong eastward jet.

![Figure 8 North-south profile of RK (\(\beta - \frac{\partial \bar{U}}{\partial y} y^2\)) for averaged east-west velocity during 10 years in the western part of the model region at (a) year 181 for Re = 209, (b) year 189 for Re = 209, (c) year 187 for Re = 314, and (d) year 188 for Re = 314.](image-url)
176 for Re = 314, and (a) year 185 for Re = 314. The corresponding periods are indicated in Figs. 3i and 3h.

Fig. 8 shows a latitude profile of RK in high and low energy periods indicated in Fig. 3h and 3i for Re = 209 and 314, respectively. In high energy periods, the sign of RK changes around 34°N and 36°N for both Re = 209 (Fig. 8a) and around 34°N for Re = 314 (Fig. 8c). That is, shear instability can potentially occur in both cases. In low energy periods, for Re = 209 (Fig. 8b), the change of the sign of RK has the same tendency as that in high energy periods, although the peak value of RK decreases slightly. On the other hand, for Re = 314 (Fig. 8d), the sign of RK is always positive and the peak value of RK becomes almost zero. That is, it is considered that the energy change for Re = 209 is not due to release of shear instability, for Re = 314 is mainly due to release of shear instability. However, it is not clear whether the energy change for Re = 314 is only due to a simple shear instability. This point requires further study.

ACKNOWLEDGEMENTS
The authors thank Dr. Masafumi Kamachi of the Meteorological Research Institute, Japan Meteorological Agency who provided the data from MOVE/MRI.COM-WNP.

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