Frontal Waves in the East of the Tsugaru Strait Revealed by the High-Frequency Radar Observation

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Abstract

Surface velocity observations of the eastern part of the Tsugaru Strait made by the high-frequency radar revealed frequent occurrence of frontal waves along the axis of the Tsugaru Warm Current in 2017–2019. The current axis (maximum of the zonal velocity in the meridional direction) disturbed in the north–south direction with period of ~13.7 days that is dominant timescale of tide modulation in the strait, in addition to that of ~27.3 days. The amplitude of the axis fluctuation increased in the downstream direction, from the eastern neck of the channel (~141.0°E) to the outlet of the strait adjacent to the Pacific Ocean (~141.5°E). The propagation speed of the disturbance was slower than that due to surface advection especially in the seasons when the stratification was developed, and agreed well with that estimated from the theory based on the two-layer baroclinic instability model except for winter. The north–south modulation of the axis at the outlet of the strait (~141.5°E) could cause short-term (from 20 days to 1 month) variations of an anticyclonic gyre of the Tsugaru Warm Current that is developed in the east of the outlet from summer to autumn reported by the previous studies.
Keywords

Tsugaru Strait, Tsugaru Warm Current, High frequency radar, Frontal waves, Tide, Baroclinic instability

1 Introduction

The Tsugaru Warm Current (TWC) is a remarkable flow in the Tsugaru Strait, between the mainland of Japan and Hokkaido, from west to east (Fig. 1a). The origin of the TWC is the Kuroshio which bifurcates in the south of the Kyushu Island, as well as the flow from the Taiwan Strait (Isobe, 1999). The current called as the Tsushima Warm Current flows northeastward along the mainland of Japan, and again the current bifurcates at the west of the Tsugaru Strait into the strait (Fig. 1a). The other branch flows northward along the Hokkaido to the Soya Strait (Fig. 1a). Kida et al. (2020) suggested that long-term trend of the volume transport of the throughflow at the Tsushima Strait is deeply related with the trend of the latitude of the eastward Kuroshio position. Thus, the variation of the TWC would be closely connected with those of these flows including the Kuroshio and the Tsushima Warm Current as one flow system (Kida et al., 2016; Kida et al., 2020).

In the Tsugaru Strait, recent studies reported significant disturbance due to the intense flow and tides (e.g., Yamaguchi et al., 2020; Tanaka et al., 2021), and subsequent characteristic
change in ocean environment there (Wakita et al., 2021). Yamaguchi et al. (2020) suggested
generation of internal waves that have large-amplitude in the vertical direction behind a
characteristic topography (sill) in the western part of the strait based on observations and
numerical experiment. Around the sill where Yamaguchi et al. (2020) made the investigation,
Tanaka et al. (2021) observed far stronger turbulence than that in the open ocean (the turbulent
energy dissipation rate: $O(10^{-5})$ W kg$^{-1}$, the vertical diffusive coefficient: $O(10^{-2})$ m$^2$ s$^{-1}$), and
estimated subsequent vertical transport of nitrate behind the sill (1 mmol m$^{-2}$ day$^{-1}$). The value
is far larger than that observed by Kaneko et al. (2021) in the subarctic region of the Western
North Pacific during summer as $\sim 1 \times 10^{-2}$ mmol m$^{-2}$ day$^{-1}$. In addition, Wakita et al. (2021)
found faster acidification in the strait based on the resent observation (2012–2019; the decrease
rate of pH is 0.0030–0.0051 year$^{-1}$) than that in the open ocean (0.0013–0.0024 year$^{-1}$; Dore et
al., 2009; Astor et al., 2013; Wakita et al., 2017; Ono et al., 2019). Moreover, Wagawa et al.
(2015) suggested that the volume transport of the TWC could affect the temperature in the
downstream coastal region, which have a big impact on aquacultures such as a kind of seaweed
(Undaria pinnatifida). Therefore, behaviors of the TWC in the strait would be subject to
investigate, because these characteristic phenomena in the strait could affect a wider region via
the TWC.

The mean velocity of the TWC in the strait reaches to 1 m s$^{-1}$ at the axis of the current
(e.g., Ito et al. 2003; Matsuura et al., 2007; Saitoh et al., 2008). Its mean volume transport was estimated as about 1.5 Sv ($10^6$ m$^3$ s$^{-1}$; Onishi and Ohtani, 1997; Ito et al., 2003) with seasonal variability of about ~0.3 Sv, showing an increase from summer to autumn (Nishida et al., 2003). The outflow from the channel encounters low temperature and salinity water described as the coastal Oyashio water by the previous studies (e.g., Kono et al., 2004; Rosa et al., 2007; Saitoh et al., 2008; Kuroda et al. 2012; Wakita et al., 2021) in the outlet of the eastern neck of the channel (~141.2°E; Fig. 1b). Thus, steep front of water mass in the north–south direction can occur in the east part of the strait (e.g., Matsuura et al., 2007; Saitoh et al. 2008; Kuroda et al. 2012), where the north boundary is open. More recent studies based on the surface velocity observation obtained from the high-frequency radar (HFR) reported seasonal variation of the latitude of the current axis; northward in summer and autumn, and sifts southward in winter and spring (Abe et al., 2020; Yasui et al., submitted to Journal of Oceanography as “Seasonal pathways of Tsugaru Warm Current revealed by high-frequency ocean radar”, hereafter Y21, and Kaneko et al., submitted to GRL as “The role of an intense jet in the Tsugaru Strait in the formation of the outflow gyre revealed using high-frequency radar data”, hereafter K21). The outflow pattern in the east of the strait (east of ~141.5°E) also changed drastically in summer and autumn, that is, large anticyclonic gyre develops south of Hokkaido (e.g., Conlon, 1982; Kawasaki and Sugimoto, 1984; Kubokawa, 1991; Rosa et al. 2007; K21).
In addition to these seasonal changes, shorter variation of TWC was also reported. Where the growth of the gyre takes for about 3 months (Yasuda et al., 1988; Nof and Pichevin, 2001), Yasuda et al. (1988) reported shorter-term variation of the gyre that has 20–30 days period, that is, the direction of the major axis of the elliptical distortion of the gyre rotated clockwise with these fluctuations. As another short-term variation in the strait, based on a mooring observation for about a year located at the southeastern corner of the channel (41.77°N, 141.12°E), Tanno et al. (2005) reported existence of remarkable periodic variation of the surface current with about 13.66 days period in addition to the four major tidal constituents (M2, S2, K1, and O1). They concluded that the variation of 13.66 days would be related with Mf tide, that is dominant timescale in the strait (e.g., Onishi et al., 2004) as well as in the Soya Strait (Ebuchi et al., 2009), and/or spring–neap tide for K1 and O1 constituents. Based on a numerical model, Kubokawa (1991) demonstrated that variation of the north–south position of the front at the outlet of the strait could affect rotation of the gyre in a clockwise direction that is resemble with the short-term variation reported by Yasuda et al. (1988), however, the connection of such short-term variation of the gyre and tides are not fully understood, yet.

Since remarkable front of the water mass and periodical disturbance due to tide are known in the strait as mentioned above, frontal disturbance related with the tide would be likely to occur in the east part of the strait (from ~141.0 to ~141.6°E, Fig. 1b). Frontal waves that are
propagating meander along remarkable front to the downstream direction were reported in other regions including the Kuroshio in the East China Sea (Sugimoto et al., 1988; Qiu et al., 1990; James et al., 1999), the Kuroshio off the Cape Shiono Misaki (Fig. 1a, Kimura and Sugimoto, 1993), the Kuroshio south of the Boso Peninsula (Fig. 1a, Itoh and Sugimoto, 2008), the Kuroshio Extension (Kouketsu et al., 2005; 2007), and the Gulf Stream in the upstream region of the Cape Hatteras (Brooks and Bane, 1981; Lee and Atkinson, 1983). Such waves frequently develop their amplitude, then developed tongue-like meander eventually detached from the main stream of the current, entraining warm-lighter (cold-denser) water in the south (north) of the front as anticyclonic (cyclonic) eddy (e.g., Yanagi et al., 1998). In addition to near surface, at the mid-depths, such frontal disturbance could contribute to water-exchange. Kouketsu et al. (2005) reported intrusion of low-salinity water into the subtropical gyre at the density of the North Pacific Intermediate Water (26.6–26.9 \( \sigma_t \); 300–600 dbar) associated with wave-like structure with the horizontal length-scale of about 100 km along the Kuroshio Extension. Therefore, if similar wave exhibits in the strait, such disturbance would be thought to play an important role for water exchange across the front, and may contribute to faster acidification as mentioned above (Wakita et al., 2021) through stirring and mixing of low pH water, in addition to significant vertical mixing that pumps up materials such as nitrate (Tanaka et al., 2021) including iron (Saitoh et al., 2008). However, observations that have resolution for such frontal
waves in the strait are rare although many repeated observations of ships and moorings were conducted there, because of the intense velocity of the TWC and frequent ship traffic through the strait.

HFR is one of the effective instruments for investigation of the frontal disturbances including eddies (e.g., Schaeffer et al., 2017). Thus, employing the HFR that includes three antennas installed at the eastern part of the strait (Fig. 1b), we investigated behavior of frontal wave through examination of the axis variation of the TWC. In the present study, the time-series of the surface velocity for three years provided by the HFR, that has horizontal resolution of about 3 km and is temporal resolution of about 30 minutes, was employed. In this paper, as a result, it will be revealed downstream propagation of the disturbance of the axis latitude with increase of the amplitude to the downstream direction. Also, it will be shown that the propagation speed is slower than that of advection due to surface current especially in the seasons of the stratification. Internal stratification and velocity distribution obtained from repeated shipboard observations along the fixed lines across the strait (Shiriya–Esan line, Fig. 1b) will be used for the investigation of the frontal waves in order to compare with the baroclinic instability theory.

The remainder of the present study is as follows. Following section, we will provide information of observation data and methods including surface velocity observed by the HFR,
shipboard observations, and propagation speed of the baroclinic instability waves based on the
theories provide by the previous studies. In the result section, first we will demonstrate the
seasonal variation of the TWC axis and shorter-term variations. Moreover, vertical distribution
of the stratification and velocity would be shown along the transection across the channel. Then,
propagation speed will be examined mainly based on the two-layer model concerning baroclinic
instability proposed by previous studies (e.g., Pedlosky, 1987; Itoh and Sugimoto, 2008). In the
final section, importance of the frontal wave in the strait will be discussed concerning short-
term variation of the gyre that develops from summer to autumn, and water mass modification.

2 Methods

2.1 High Frequency Radar Velocity

In order to monitor the surface current in the eastern part of the Tsugaru Strait, the
Mutsu Institute for Oceanography Ocean (MIO), Japan Agency for Marine-Earth Science and
Tecnology (JAMSTEC) has installed the monitoring system using the HFR (CODAR,
SeaSonde, 13.9 MHz) that has three antenna stations (Fig. 1c). Two of them are installed in the
Shimokita Peninsula, and the other is located at Hokkaido (Esan, Fig. 1c). The HFR provides
data whose horizontal resolution of the system is about 3 km, and the coverage of the
observation ranges from about 3 km to 60 km from each antenna. The surface current
distribution is calculated almost each 30 minutes by the system and uploaded at the site of MIO Ocean Radar data Site for the eastern Tsugaru Strait (MORSETS; http://www.godac.jamstec.go.jp/morsets/e/top/). Abe et al. (2020) reported that the root-mean-square error between the HFR surface velocity and the sub-surface velocity obtained by acoustic Doppler current profiler mounted on some ships (SADCP) in the Tsugaru Strait was 30 (26) \(\times 10^{-2}\) m s\(^{-1}\) in the east–west (north–south) direction. K21 suggested a similar accuracy of the HFR velocity with that of Abe et al. (2020), while calling attention to an effect of the surface wind on the HFR velocity especially during winter. The same HFR is used for the surface current monitoring at the Tsushima Strait and at the Soya Strait (Fig. 1a), and some previous studies have been demonstrated using them; the Tsushima Strait: Yoshikawa et al. (2006; 2010), and the Soya Strait: Ebuchi et al. (2009). Yoshikawa et al. (2006) mentioned that the root-mean-square velocity difference between the HFR and SADCP was 6.62–11.3 \(\times 10^{-2}\) m s\(^{-1}\). Ebuchi et al. (2009) reported that the current estimated from the HFR agreed well with those derived from SADCP with root-mean-square difference less than 25 \(\times 10^{-2}\) m s\(^{-1}\). In the present study, we employed data from 2017 to 2019 when the monitoring was stably conducted by the all three antennas (the monitoring itself started since 2014).
2.2 Analysis for the Temporal Variation of the Axis of the Tsugaru Warm Current

First, we defined twenty-five sub-regions that is long in the north–south direction (Fig. 1c; R1–R25), and made spatial mean of zonal velocity in the east–west direction in the subregion. Further, temporal mean of each 6-hour concerning the spatial mean was calculated in each sub-region, to obtain the north–south distribution of the zonal velocity along each sub-region. Then, the latitude where mean zonal velocity showed the maximum was defined as the axis of the TWC. We investigated the seasonal mean location of the axis as well as its standard deviation (Figs. 2 and 3). We also focused shorter-term variation of the axis (some examples are shown in Fig. 4), and calculated spectrum of the temporal variation of the latitude of the axis in each sub-grid, using data from 2018 to 2019 (Fig. 5; there was no missing of the HFR data during the duration, but 2017). Moreover, we obtained low-pass-filtered data using a 5th butterworth filter that has the cutoff timescale of 10-days (Fig. S1) in order to focus on variations from 10 days to one month (Yasuda et al., 1988), for each three months. Then, we made lag-correlation analysis of the time-series data with the low-pass concerning each sub-region in relation to that in the reference sub-region near the eastern neck of the channel (R1; 140.96–140.04°E; Fig. 1c), expected as the region where disturbance of the axis would be generated (Fig. 6). The timescale of the correlation peaks was regarded as the lag of arrival of the disturbance, and using the distance between the reference sub-region and each sub-region,
propagation speed of the disturbance was estimated in each season (winter: January–March, spring: April–June, summer: July–September, autumn: October–December) for each year (Fig. 7).

2.3 Shipboard Observations

Repeated observations were conducted along the SE-line between Shiriya and Esan (Fig. 1b, Table 1) in almost each four seasons from 2009 to 2019 (Table 2) by three ships; the training ship (T/S) Ushio-Maru and T/S Oshoro-Maru belonging to Hokkaido University, and the research vessel (R/V) Wakataka-Maru belonging to the Japan Fisheries Research and Education Agency.

Observations of temperature, salinity, and pressure were made as conductivity-temperature-depth (CTD) observation in the cruises, using SBE 911 plus (Sea-Bird Scientific, Inc.). Following the algorithms of UNESCO (United Nations Educational, Scientific and Cultural Organization) (1983), depth was calculated from the value of the pressure and the latitude of the stations. Then, each 1 m mean of temperature, salinity, and pressure was estimated using linear interpolation. The potential density anomaly, $\sigma_0$, was calculated using the temperature, salinity, and pressure of each 1 m following Gill (1982). Then, an isodepth mean of $\sigma_0$ in each season was calculated again (Fig. 8). After that, geostrophic velocity was
calculated following Pond and Pickard (1986) from each seasonal isodepth mean properties
between the adjacent stations (Fig. 9). The reference level was set to the bottom.

2.3 Theoretical Estimation of Propagation Speed of the Baroclinic Instability in
the f-plane

Since the meridional width of the eastern part of the Tsugaru Strait is narrow (∼0.4
degree of latitude around the SE-line although north boundary is open, Fig. 1b) and thus impact
of latitudinal difference in the inertial frequency on the phenomena would be small. For this
reason, f-plain approximation would be adequate; the difference of the inertial frequency
between at the northern boundary of the strait ∼41.8°N and the southern boundary ∼41.4°N is
<1 %. On the other hand, vertical structure is known to change drastically in each season (e.g.,
Sugimoto and Kawasaki, 1984). Therefore, it should be treated with caution for assumption of
the baroclinicity. In summer, intense pycnocline was reported (e.g., Matsuura et al., 2007;
Saitoh et al., 2008), which implies an applicability of the two-layer model (e.g., Pedlosky, 1987).
On the other hand, in winter, it is well known that surface mixed layer is developed and density
is vertically homogeneous (e.g., Sugimoto and Kawasaki, 1984). Thus, it is expected that
assumption of baroclinicity probably does not match well in winter, but we included winter in
the present investigation for comparison. In spring and autumn, stratification would be weaker
than that in summer. For this reason, we also employed the $f$-plain model that has an assumption of continuous stratification as well as constant vertical shear (Eady, 1949).

Using isodepth mean of the seasonal mean density anomaly from station SE3 to SE7, $\langle \sigma_\theta \rangle_{SE}$, we defined the depth of the boundary between the upper- and lower-layer as follows:

$$\sigma_r = \sigma_2 - (\sigma_2 - \sigma_1)/e, \quad (1)$$

where $\sigma_1$ and $\sigma_2$ were potential density anomaly at 10 m and 250 m of $\langle \sigma_\theta \rangle_{SE}$, respectively. The depth has the potential density anomaly of $\sigma_r$ was defined as $H_1$, that is the upper-layer thickness (Fig. 10a). Then, following the derivation of Pedlosky (1987) (the equation 7.11.9), the phase speed of the baroclinic instability in the case of zero-planetary beta ($\beta=0$) in the two-layer stratified model as follows:

$$c = \frac{U_1(K^2+2/R_1^2)+U_2(K^2+2/R_2^2)}{2(K^2+1/R_1^2+1/R_2^2)} \pm \frac{-(U_1-U_2)^2(4/R_1^2R_2^2-K^4))^{1/2}}{2(K^2+1/R_1^2+1/R_2^2)} \quad (2)$$

where $K$, $U_1$, and $U_2$ are the total wavenumber, the mean velocity of the upper-, and that of the lower-layers, respectively. And

$$\frac{1}{R_n^2} = \frac{f_{41.5^oN^2}}{g'H_n}, \quad (3)$$

where $H_n$ is the thickness of the n-th layer, and $g'$ is the reduced gravity calculated as follows:

$$g' = \frac{(\rho_2-\rho_1)}{\rho_0} g, \quad (4)$$

where $g$ is the gravitational acceleration (we employed it as 9.8 m s$^{-2}$), and $\rho_1$, and $\rho_2$, are mean density in the upper- and lower-layer, respectively. We defined the reference density, $\rho_0$, as 1026
We employed the inertial frequency at 41.5°N \( (f_{41.5°N} \approx 18.1 \text{ hour}) \).

The instability occurs when \( K^2 < 2/R_1R_2 \). Assuming the wavenumber of the disturbance observed in the strait would be small sufficiently for such instability, we employed the propagation speed of the instability (real part of the Eq. 2), as mentioned in Itoh and Sugimoto (2008), as follows:

\[
c = \frac{U_1(K^2+2/R_2^2)+U_2(K^2+2/R_1^2)}{2(K^2+1/R_1^2+1/R_2^2)}.
\]

(5)

The upper-layer thickness, \( H_1 \), was calculated as mentioned above, and the lower-layer thickness was defined as the difference from 300 m minus \( H_1 \), assuming that typical water depth in the eastern part of the Tsugaru Strait beneath the axis is about 300 m (Fig. 1b). Then, \( U_1 (U_2) \) was estimated using the spatial mean the geostrophic velocity in the upper-layer (lower-layer) along the SE-line (generally stations SE3–SE7, but SE4–6 in winter; Fig. 10b) in each season.

When the horizontal wave number \( K \) is sufficiently smaller than \( 1/R_n \), in other words, the wavelength of the frontal wave is far longer than the deformation radii, the Eq. 5 can be further simplified as follows:

\[
c = \frac{U_1(1/R_2^2)+U_2(1/R_1^2)}{(1/R_1^2+1/R_2^2)}.
\]

(6)

This equation indicates that the propagation speed, \( c \), takes a value between \( U_1 \) and \( U_2 \) with
$1/R_n^2$, as the weighting function.

In addition to the two-layer model, for comparison, we calculated the phase speed of the baroclinic instability in the continuous stratification and constant vertical shear of zonal velocity following Eady’s theory as follows:

$$c_E = \frac{u_s}{2} \pm \frac{u_s}{\mu} \sqrt{\left(\frac{\mu}{2} - \cosh\frac{\mu}{2}\right) \left(\frac{\mu}{2} - \tanh\frac{\mu}{2}\right)},$$  \hspace{1cm} (7)

where $u_s$ is the surface velocity, and $\mu = L_D K$. $L_D$ is deformation radius defined by $L_D = NH f_0$ ($N$ is Brunt–Väisälä frequency in the continuous stratification as mentioned later; $H$ and $f_0$ are layer thickness and inertial frequency at the reference latitude, respectively). Note that considering occurrence of the instability, we defined the propagation speed in the Eady’s case as a real part of the Eq. 7, that is $c_E = u_s/2$. Here, $u_s$ was calculated as a mean geostrophic velocity for upper 10 m.

It should be also noted that in order to determine the wavelength of the disturbance in each season, we employed the maximum growth rate of Eady’s case, $\mu = 1.61$. We calculated $N$ as follows:

$$N = \frac{1}{H} \int \sqrt{-\frac{g}{\rho_0} \frac{\partial (\sigma \theta) SE}{\partial z}} \, dz. \hspace{1cm} (8)$$

For estimation of $L_D$, $H$ and $f$ were set as 300 m and $f_{41.5^\circ N}$, respectively. Using these equations and parameters, we calculated the propagation speed of the baroclinic instability in some cases, and compared them to the propagation speed of the observed disturbance of the TWC’s axis in
3 Results

3.1 Frontal waves detected from the surface current distribution observed by HFR

First, seasonal distribution of the axis was shown briefly, before demonstration of the short-term variations of the axis. Fig. 2 showed seasonal mean location of the axis in each sub-region together with standard deviation. From the eastern neck of the channel (~141.0°E) to the Shiriya Spur (~141.5°E), whereas the axis located relatively south (south of 41.6°N) during winter and spring (Fig. 2a and b, respectively), it moves northward from summer to autumn (north of 41.6°N, Fig. 2c and d, respectively). This seasonal movement has been already reported by the previous studies (e.g., Rosa et al., 2007; Abe et al., 2020; Y21; K21). The velocity of the axis was generally faster (slower) in summer (winter) reaching 1.0 (0.6) m s\(^{-1}\) west of the Shiriya Spur. With respect to the east of the spur, the velocity was slow in winter and spring (Fig. 2a and b, respectively). In contrast, relatively larger velocity and somewhat straightforward flow pattern in the east–west direction was observed in summer (Fig. 2c). Standard deviation of the axis latitude generally increased to the downstream especially in the region of 141.0–141.6°E, that is west to the Shiriya Spur (Fig. 3). As an exception, in the season
of winter the standard deviation showed a small peak at ~141.4°E, at the west-side of the Shira Spur (Fig. 3a). In spring, the standard deviation increased monotonically with increase of the longitude in the region west of 141.8°E (Fig. 3b), while in summer, the standard deviation increased steeply in the range of 141.4–141.6°E (Fig. 3c). Concerning autumn, the distribution of the standard deviation in October (December) was similar with that in spring (winter) (Fig. 3d).

Besides the seasonal movement of the axis as mentioned above, shorter-term oscillation within several dozen days was observed. We showed some examples of such variation as Fig. 4. In summer of 2017, large amplitude oscillation of the axis in the north–south direction was recognized especially north the Shiriya Spur as ridge-shaped distribution that has one north convex peak with the east–west scale of ~50 km (Fig. 4a). On the other hand, in winter of 2018, short-term variations were also active (Fig. 4b), but multiple peaks of the meander were demonstrated, and the horizontal scale in the east–west direction of the wave-like meandering (peak-to-peak) was inferred as ~40 km. The short-term variations were also observed in the other seasons (not shown).

We estimated power spectral density concerning the oscillation of the axis in each sub-region using data from 2018 (January 1st) to 2019 (December 31st) when the surface velocity continuously observed (Fig. 5). The spectrum indicated remarkable peaks around timescale of
13.66 days as well as that of double of it (27.32 days) in the western regions from the eastern neck of the channel (R1) to the Shiriya Spur (around R12; Fig. 1b and c). Peaks of the timescale of ~13.7 days and ~27.3 days became obscured in the regions east to the spur. The variation with ~13.7 days was consistent with that reported near the eastern neck of the channel (around Esan, Fig. 1b) by Tanno et al. (2005).

In order to examine the relationship of the periodical disturbance among each sub-region, we defined the sub-region R1 (Fig. 1c) as the reference where the disturbance was expected to be generated, and then, we calculated lag correlation of the time-series of the axis latitude between the reference sub-region and other sub-regions (R2–R25) (Fig. 6). In order to focus the variation that has longer timescale than 10 days (Yasuda, 1988), we calculated the low-pass-filtered data using the fifth power butterworth filter that has the cutoff timescale of 10-days (Fig. S1). Significant positive peak was recognized in R2–16, showing increase of the lag with the increase of the distance from the reference sub-region (Fig. 6). The lag peaks showed a cyclical characteristic in each sub-region with a cycle of ~14 days (Fig. 6).

We plotted the lag-time of the correlation peaks (shown by triangles in Fig. 6) concerning the positive range (y-axis), against the distance from the reference sub-region (x-axis) (Fig. 7). Here, in addition to the first cycle, the lag of the second cycle starting at 13.7 days, is also shown as the gray plot. The lag-correlation was calculated in four seasons (3
months) of each year (2017–2019). We also estimated the timescale of advection estimated from the mean surface velocity $U_1$ from R1 to R25 (gray broken lines with smaller slope in Fig. 7). Linear increasing of lag-time of the observed disturbance with increase of the distance (symbol plots in Fig. 7) was clearly demonstrated in spring and summer. The slopes of the disturbance (symbol plots in Fig. 7) indicated seasonal variation, showing the largest angle in spring. We regarded the slope as the propagating speed of the propagation of the axis meandering. That is, it was suggested that the propagation speed of the observed disturbance became slowest in spring. In winter, propagation of the disturbance of the TWC’s axis was unclear in 2017 and 2018 (Fig. 7d). To sum up, the results showed that the longer timescale concerning the propagation of the disturbance was frequently recognized than that estimated from the surface advection, especially in the seasons from spring to autumn in the regions west of the SE-line (Fig. 7a–c).

The slower propagation speed than that of advection suggested the disturbance of the axis would not be simply advected by the surface current. Thus, we suspected the propagation of the disturbance of the TWC’s axis as propagation of the frontal waves as reported in other regions (e.g., Brooks and Bane, 1981; Lee and Atkinson, 1983; Sugimoto et al., 1988; Kimura and Sugimoto, 1993; Kouketsu et al., 2005; Itoh and Sugimoto, 2008), as mentioned in the Introduction. Since meridional range of the frontal disturbance in the strait is relatively smaller
than those reported by the previous studies in the western boundaries, thus, it is easy to think that the $f$-plain assumption seems to be reasonable. Whereas, structure of stratification in the strait was known to be drastically changed in each season as mentioned above (e.g., Nishida et al. 2003; Saitoh et al., 2008). Thus, to examine the application of the two-layer baroclinic instability theory, we would demonstrate the seasonal vertical structure across the strait concerning the density and velocity in the next subsection.

3.2 Internal structure concerning the stratification and velocity across the channel obtained from shipboard observations

The HFR revealed near surface disturbance that had slower propagation speed than that caused by advection of the surface current especially in stratified season as mentioned in the previous subsection. Such downstream-propagating disturbance with increasing amplitude might be expected as baroclinic instability along front. To test the theory of the instability under the two-layer model condition as described in the Material and Methods section, internal structure including the upper- and lower-layer thickness and velocity there, are required (Eq. 5). Thus, in this subsection, we would show seasonal distribution of the thickness and velocity across the eastern part of the strait (SE-line; Fig. 1b) obtained from shipboard observations of 2009–2019 (Table 2).
Seasonal mean of the salinity showed that higher salinity water located the southern side of the strait (Fig. 8), indicating a remarkable front around the center of the strait (~41.7°N except for winter; ~41.6°N in winter). The location of the front corresponded well to that of the axis of the TWC (Fig. 2). Moreover, this salinity distribution suggested that high-temperature and high-salinity water of the TWC was distributed along the southern coast, while lower-temperature and lower-salinity water affected by the coastal Oyashio was located in the northern side of the strait as reported by the previous studies (Kono et al., 2004; Rosa et al., 2007; Saitoh et al., 2008; Kuroda et al. 2012; Wakita et al., 2021). The level of the salinity at the core of the water of the TWC was highest in summer (Fig. 8b). In autumn, salinity contrast in the north–south direction weakened showing a general increase of the salinity (>34.8 psu) in the strait (Fig. 8c).

Also, seasonal mean transection along the SE-line of the potential density anomaly, $\sigma_\theta$, indicated remarkable variation, showing strong (weak) stratification in summer (winter) (Fig. 8). In winter, mean $\sigma_\theta$ in the channel were estimated as ~26.5 $\sigma_\theta$ and vertically almost homogeneous. In contrast, steep surface pycnocline at the subsurface, ~50 m, was estimated as a mean distribution in summer, showing that the surface (near bottom) potential density anomaly was ~23.0 (~26.0) $\sigma_\theta$. In spring and autumn, although vertical difference of the density was weaker than that in summer, stratification was also developed.
Similar as the potential density anomaly distribution, geostrophic velocity distribution along the same transection also showed remarkable seasonal variation (Fig. 9). In winter, the velocity distribution was expected to be vertically homogeneous and the magnitude of the eastward flow was weak. In the other seasons when stratification was developed, remarkable eastward current core (>1 m s$^{-1}$) was calculated near surface. In summer and autumn, opposite westward flow was estimated south of the intense eastward current in the upper-layer. It should be noted here that the estimation of eastward geostrophic velocity in autumn might be overestimate due to the westward bottom current near the bottom around stations 3–4 that was suggested by the SADCP observation (not shown).

Seasonal distribution of $\sigma_0$ and velocity suggested the obvious stratified situation in the strait except for winter. This result was consistent with that reported by many previous studies (e.g., Sugimoto and Kawasaki, 1984; Nishida et al., 2003; Matsuura et al., 2007; Saitoh et al., 2008). Thus, we estimated the representative values of the upper- and lower-layer thickness, and velocity in each season using the Eq. 1. The depths of the upper-layer, $H_1$, in each season were estimated as 63 m (spring), 69 m (summer), 124 m (autumn), and 152 m (winter) (denoted by the triangles in Fig. 10a). The depths, $H_1$, were indicated as magenta horizontal broken lines in Figs. 8 and 9, and the estimation seemed reasonable as the boundary of the upper- and lower-layer over the transection. Note that the remarkable two-layer like distribution was
demonstrated in summer, but in the other stratified season (spring and autumn), the vertical change in the density seemed somewhat gentle (Fig. 10a). This characteristic was similar regarding seasonal change in the vertical structure of geostrophic velocity (Fig. 10b). Thus, not only the approximation of the two-layer model, but also that of the continuous stratification was also investigated in the present study as mentioned in the next subsection (3.3), using the Eq. 7.

3.3 Investigation of the propagation speed of the frontal disturbance

Using the parameters estimated in the previous subsection, and the Eqs. 5 and 7, we calculated the propagation speed of the baroclinic instability in order to compare them to the propagation speed of the disturbance observed by the HFR (Fig. 7). We employed the wavelength of baroclinic instability of each season as 70 km (spring), 145 km (summer), 98 km (autumn), and 43 km (winter), respectively, assuming the maximum growth case in the Eady’s model. This wavelength satisfies the instability condition of the two-layer model (e.g., the equation 7.11.10 of Pedlosky, 1987).

In spring, the propagation speed of the observed disturbance was similar with that of the two-layer model (0.14 m s\(^{-1}\); Fig. 7a) in the distance range of 30–80 km, although outliers were shown in the downstream of the Shiriya Spur in 2018. In contrast, the propagation speed of the Eady’s model were somewhat slower than that of the observed disturbance. In summer,
the propagation speed of the baroclinic instability $\sim 0.20 \text{ m s}^{-1}$ was also consistent with that of the observed disturbance especially in 2019 (Fig. 7b). It should be noted that, the zonal velocity beneath the axis of the TWC was smaller in 2018 than that in 2019 according to the SADCP observation, showing a consistency with slower propagation speed in 2018 (not shown).

Concerning autumn, because the propagation of the disturbance of the axis was not clear in 40–80 km, thus it is difficult to detect clear correspondence between them (Fig. 7c). However, when a continuous propagation of the disturbance was assumed between that in 0–40 km and 80–110 km in 2017 (upward triangles in Fig. 7c), the propagation speed seemed to show a similarity with that of the two-layer model (0.28 m s$^{-1}$) and Eady’s model. In the season of winter, the propagation of the disturbance was not clear in 2017 and 2018 (Fig. 7d). In 2019, the distribution of the downward triangles is not uniform in slope on the 0–100 km scale. Moreover, the propagation speed of the two-layer model was far smaller than that of the observed disturbance because of weak eastward geostrophic velocity. In winter, stratification was very weak in the region (Fig. 10a). Thus, caution should be required in the interpretation of the propagation of the disturbance in this season as a baroclinic instability.

As a summary of this subsection, it could be said that the propagation speed in stratified season showed a well agreement with that of the baroclinic instability in the case of the two-layer model, especially in spring and summer. Therefore, the propagation of the axis disturbance
of the TWC observed by the HFR in the seasons was suggested as that closely related with the
baroclinic instability.

4 Discussion

4.1 Compatibility of the two-layer model to the observed disturbance, and
comparison with frontal waves in other regions

In the present study, we examined the consistency of the propagation speed of the
disturbance of the TWC’s axis observed by the HFR with the propagation speed based on the
theory of the two-layer baroclinic instability in the $f$-plane (Fig. 7). Then, close value of the
propagation speed of the disturbance of the TWC’s axis observed by the HFR with that
calculated assuming the two-layer model was indicated in the stratified season, such as spring
and summer in a cycle of $\sim$13.7 days (Fig. 7a, and b). In addition to the two-layer model, we
also examined the propagation speed based on the Eady’s theory. Although the magnitude of
the propagation speed based on the Eady’s theory was similar with that of the observed
disturbance in summer and autumn, however, they were dissociated in the downstream region
($>40$ km, Fig. 7a) in spring. Also, it should be noted that, considering the characteristic of the
internal structure, the two-layer assumption also seemed better than that of continuous
stratification especially in summer. The two-layer assumption may be better for the autumn
season as well. The geostrophic velocity in the lower layer in autumn may perhaps be overestimated, because some of the observed disturbances seemed slower than that of the Eady’s theory. In the SADPC observation, westward velocity was sometimes observed near the bottom between the stations 3–4 (not shown), which may be due to topographic wave propagation from the north related to the wedge-shaped topography in the eastern part of the strait (Figure 1b). Moreover, we would like to note that more careful investigations are needed for determination of the dominating wavelength in the region, since the dominant wavelength was assumed based on the Eady's theory.

The observed propagation speed in summer (0.20 m s\(^{-1}\)) was similar with that estimated by Itoh and Sugimoto (2008) near the Boso Peninsula (Fig. 1a; 0.22–0.30 m s\(^{-1}\) with the wavelength of 220–380 km), and Kouketsu et al. (2007) in the Kuroshio Extension (0.20–0.30 m s\(^{-1}\) with the wavelength of ~200 km). In contrast, the propagation speed was smaller than that reported in the Gulf Stream by Lee and Atkinson (1983) (0.50–0.70 m s\(^{-1}\) with the wavelength of 300–500 km). Although the water depth was different between the TWC and the open ocean currents such as the Kuroshio and the Kuroshio Extension, the effect of ratio concerning the thickness of the upper- to the lower-layer would possibly be accountable for this similarity. The slower propagation speed in springtime (0.14 m s\(^{-1}\)) may be due to the strength of the surface currents \(U_1 = 0.20\) m s\(^{-1}\) that was weaker than that in the summer season \(U_1 = \)
As other issues to be considered, we would like to mention the topography. An impact of the characteristic topography on the instability was not considered in the present study. The steep increase in amplitude over the Shiriya Spur in summer (Fig. 3c), and the bypassing movement of the axis around the spur (Fig. 4a) may suggest the influence of the topography. On the other hand, short-term variation of the axis also indicated a riding up of the TWC over the spur sometimes (Fig. 4a). The outliers in Fig. 7a around the spur in 2018 (with lag of ~10 days) may be also related with the topographic effect. Y21 suggested that the TWC tends to flow over the spur with increase of inertia in winter. Thus, the relationship between the topographic effect and the path of the TWC should be also investigated as an issue to address in relation to intensification of the inertia of the TWC.

4.2 Frontal Disturbance in the Weak Stratified Season

Figure 4 implies that the length-scale of the frontal disturbance in the east–west direction seemed smaller in winter than that in summer. In winter, as estimated $L_D$ was small, it might occur the scale of the instability become smaller. However, as mentioned above, the accuracy of the application of the baroclinic instability is suspicious for the homogenous density distribution (Fig. 10a). Instead, barotropic disturbance along the shelf might be plausible. In
relation with the barotropic phenomena, using idealized topography, Ohshima (1994) pointed out scattering of the Kelvin wave to the higher mode shelf waves at the eastern outlet of the strait (northeast corner of the Shimokita Peninsula) based on a numerical model. Although there is difference in the topography between the idealized one of Ohshima (1994) and reality, similar scattering might be caused by the topography in the eastern neck of the channel (~141.1°N), because the mouth of the strait becomes wider there in the eastward direction (Fig. 1b). These shelf waves might affect the short-term disturbance of the TWC’s axis along the north coast of the Shimokita Peninsula in winter (Fig. 4). Especially higher mode of the shelf waves might cause slow propagation of the disturbance of the axis (Fig. 7d, that in the range of 0–40 km of 2019). Moreover, the eastward propagation of the disturbance, which was more pronounced in winter (Fig. 7d, 2018), may imply the influence of topographic trapping waves from north to east associated with the wedge-shaped topography in the eastern part of the strait as mentioned above (Fig. 1b). Wintertime behavior of the TWC should be examined using the HFR and numerical model in future.

4.3 Impact of the Disturbance on the Short-term Variation of the Tsugaru Gyre

As mentioned in the Introduction, observational studies by Yasuda et al. (1988) have reported short-term fluctuations of the large anticyclonic gyre which timescale was 20–30 days
(the direction of the major axis of the gyre's elliptical distortion rotated clockwise with these fluctuations). Kubokawa (1991) showed in his numerical experiment that such clockwise variations along the gyre can be successfully reproduced by varying the volume transport of the TWC that has relatively lower potential vorticity by changing the latitude of the front of the TWC at the outlet of the strait in the north–south direction. Therefore, it is quite possible that the north–south frontal disturbances observed in this study could influence such short-term variations of the large anticyclonic gyre. At the eastern outlet of the strait (e.g., R12; 141.42–141.50°E), a peak of ~27.3 days was also demonstrated in the spectrum of the frontal disturbance (Fig. 5), which is consistent with the periodicity reported by Yasuda et al. (1988). On the other hand, although the period of ~13.7 days was also recognized at the outlet of the strait, the period is a little shorter than the timescale reported by Yasuda et al. (1988). Thus, it is necessary to examine whether the shorter-term variability of the gyre at this timescale also occurs, as a future work. Since the period of the frontal disturbance is in good agreement with the ~13.7 days cycle (and its double length) that prevails in the Tsugaru Strait and the Soya Strait (Onishi et al., 2004; Tanno et al. 2005; Ebuchi et al., 2009), the result suggested that tides may play an important role on the frontal disturbance. However, it is not unclear yet that what specific processes are involved between them. This is another issue that needs to be continued to be examined. The short-term variability of the large anticyclonic gyre (gyre mode) is
considered to be closely related to changes in the fishing grounds of pelagic fishes around the Shimokita Peninsula (e.g., Sato 1974, Hirai et al. 1988), and thus, the elucidation and forecasting of the excitation process is an important issue in terms of contributing to the prediction of fishing grounds. If the relationship between the tide and the short-term variability of the gyre is clarified, there is a high possibility that it will eventually contribute to the prediction of the fishing grounds.

4.3 Importance of the Baroclinic Instability on the Water modification in the Tsugaru Strait

As pointed out by the many previous studies, the development of the frontal wave could affect water mass exchange in the direction across the front (e.g., Yanagi et al., 1998; Kouketsu et al. 2005). In the region of the frontal disturbance was observed by the HFR, the low temperature and salinity water called as the coastal Oyashio Water sometimes juts from north (e.g., Kuroda et al. 2012; Wakita et al., 2021). Thus, the development of the disturbance may contribute to the transport of such cold (warm) water to the northern coast of the Shimokita Peninsula (coastal area of Hokkaido). Actually, the north–south contrast of salinity in the strait rapidly weakened in autumn (Fig. 8c), the after season of summer when the remarkable occurrence of the instability would be expected. Also, although it is a case in winter, coastal
monitoring conducted by the MIO revealed that intense negative anomaly from the annual mean (about 4 °C lower) in 2014 around the northern coast of the Shiomokita Peninsula. In this year, landing of the yellow goosefish (*Lophius litulon*) was poor, which brought large social impact around the fisheries communities along the northern coast of the peninsula. Thus, investigation of the mechanisms of such water exchange and subsequent forecast are important not only for science but also for the society.

In addition, the instability may also affect vertical mixing of the waters through some mechanisms such as local intensification of the vertical shear. Kouketsu et al. (2007) suggested that intrusion of the denser water at the mid-depths could occur in the region of the crest to the trough of the upper frontal wave in the Kuroshio Extension due to the baroclinic instability. Thus, local enhancement of the turbulence associated with such intrusions could be also expected, which might contribute to the characteristic distribution of the turbulence intensity along stream direction. The grasp of the along stream structure of the turbulence and fluxes would bring more accurate estimation of them in the frontal region as a progress study of the previous understanding such as Kaneko et al. (2012; 2013). The observations of the turbulence and fluxes including low pH water will be conducted in near future. Moreover, these studies of water mass exchanges and vertical mixing (including the present study) would also contribute to improve understanding of the mechanism of the faster acidification the Tsugaru Strait than
that in the open ocean (Wakita et al., 2021).

5 Conclusions

This study focused on the front, that is the current axis, in the eastern part of the Tsugaru Strait where the low-temperature, low-salinity water of the Oyashio system and the high-temperature, high-salinity water of the Tsugaru Warm Current meets, using data provided by the HFR that enables quasi-real time monitoring of a wide area, and the CTD ship data accumulated over 10 years. The north–south disturbance of the axis propagated eastward (downstream) with increasing amplitude, especially during the stratification period, with a periodicity of about 14 days, which is the predominant period of tide in this region. This result suggested a possibility of baroclinic instability near the front in this region. These disturbances would affect the north–south water mass mixing in the strait and the short-term variability of the Tsugaru warm water gyre that develops on the Pacific side during the stratification period.

In conclusion, this study suggests that tidal fluctuations in the Tsugaru Strait could cause short-term changes in the gyre of the TWC outflow through the propagation along the water mass front, contributing to the understanding of the individual findings as an integrated phenomenon.
**Abbreviations**

CTD: conductivity-temperature-depth (observation); HFR: High Frequency Radar; JAMSTEC: Japan Agency for Marine-Earth Science and Technology; K21: Kaneko et al., submitted to GRL as “The role of an intense jet in the Tsugaru Strait in the formation of the outflow gyre revealed using high-frequency radar data”; MIO: Mutsu Institute for Oceanography Ocean; MORSETS: MIO Ocean Radar data Site for the eastern Tsugaru Strait; R/V: research vessel; SADCP: acoustic Doppler current profiler mounted on some ships; SE: Shiriya–Esan (line); T/S: training ship; TWC: Tsugaru Warm Current; UNESCO (United Nations Educational, Scientific and Cultural Organization); Warm Current revealed by high-frequency ocean radar”; Y21: Yasui et al., submitted to Journal of Oceanography as “Seasonal pathways of Tsugaru

**Declarations**

**Availability of data and material**

The datasets of surface velocity obtained from the HFR analyzed for this study can be distributed through the Mutsu Institute for Oceanography, JAMSTEC (https://www.godac.jamstec.go.jp/morsets/e/top/). The other data analyzed for this study are available from the corresponding author upon reasonable request.
Competing interests

The authors declare that they have no competing interest.

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Authors' contributions

HK conceptualized the present study, and performed analysis of the data obtained from the HFR and shipboard observations, visualization of figures, and writing the manuscript reflecting the comments from all authors. HK, TT, HA, MW, KS, DH, and TO devised the observation plan and engaged the in-situ data collection together with ST. SW and KS projected an introducing and installing the HFR, and they also administrated and supervised concerning the continuous observation of the HFR. YS was responsible for the management and retention of the HFR data. All authors were contributed to discussions of the results and comments on the manuscript.

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**Figure legends**

Figure 1. (a) Schematic diagram of the flows around Japan. (b) Locations of stations along the repeated observation lines around the Tsugaru Strait. (c) Observation site of the present study and distribution of the temporal mean velocity at each grid point for 2017–2019. Contours in (b) denote topography. The square grids in (c) indicate the bins used to calculate the zonal average of the zonal velocity.

Figure 2. Mean latitude of the axis of the TWC (squares) with standard deviation (blue lines) at each grid in each season of (a) winter, (b) spring, (c) summer, and (d) autumn. Colors of the
squares denote mean value of the zonal velocity at the axis. Contours and triangles are same with those in Fig. 1.

Figure 3. Standard deviation of the axis of the TWC calculated using the bootstrap method (thick lines) in each monthly of (a) winter, (b) spring, (c) summer, and (d) autumn. Thin lines indicate 95% confidence interval of the bootstrap method. Broken lines correspond to the location of the Shiriya Spur.

Figure 4. Examples of short-variation of the axis of the TWC. Each-one-day-mean of the latitude of the axis of the TWC is shown. The upper row labeled as (a) is that in August in 2017. The lower row labeled as (b) is that in February in 2018. Contours and triangles are same with those in Fig. 1.

Figure 5. Power spectral density (PSD) of the latitude of the TWC’s axis calculated from data for 2018–2019 in each sub-region of (a) R1–R12, and (b) R13–25. Lat, and CPD mean the degree of latitude, and cycle per day, respectively. For ease of viewing, each data is displayed with slightly shift of increments in the vertical axis direction. Thin solid lines denote 95% confidence interval. Dark (light) gray broken line denotes the timescale of 13.66 (27.32) days.

Figure 6. Examples of lag-correlation ($r$) of some sub-regions in relation to the reference sub-region (R1) in 2017. Triangles show peaks of the lag correlation. Thin lines denote 95% confidence interval.
Figure 7. Distance–lag plot in each season; (a) spring, (b) summer, (c) autumn, and (d) winter. Plots with dark(light)-gray-color denote the first (second) cycle of the propagation starting from the lag 0 (13.7) days. Thin (thick) solid line denotes the propagation speed estimated from the Eq. 4 (based on Eady’s theory). Broken lines indicate advection time-scale of the upper- and lower-layer. Vertical solid (broken) lines denote the corresponding location of the Shiriya Spur (Shiriya–Esan-line).

Figure 8. Vertical transection of seasonal mean of the salinity (color), and potential density anomaly, $\sigma_0$ (contour), along the Shiriya–Esan-line in (a) spring, (b) summer, (c) autumn, and (d) winter. Triangles show the latitude of each station. Circle with error bar is a position of the TWC axis for 2017–2019 with standard deviation. Broken magenta line denotes the depth of the upper-layer, $H_1$.

Figure 9. As Fig. 8, but for the geostrophic velocity across the Shiriya–Esan-line (outflow from the strait is positive). (a) spring, (b) summer, (c) autumn, and (d) winter.

Figure 10. Typical vertical structure in each season of (a) potential density anomaly, $\sigma_0$, and (b) zonal velocity in the Esan–Shiriya-line. Note that whereas (a) is the seasonal isodepth mean of $\sigma_0$, from the station SE3 to SE7, (b) is the seasonal isodepth mean of the geostrophic velocity for the stations SE3–SE7 (as an exception, SE4–SE6 in winter). Triangles are the upper-layer depth, $H_1$, in each season.
Fig. S1. Example of low-pass-filtered timeseries of the latitude of the TWC’s axis in 2017, using the fifth-order butterworth filter that has the cutoff timescale of 10 days.
## Tables

Table 1. Locations of the stations of the repeated shipboard observations.

| Station | Latitude (°N) | Longitude (°E) |
|---------|---------------|----------------|
| SE1     | 41.44         | 141.42         |
| SE2     | 41.48         | 141.40         |
| SE3     | 41.53         | 141.37         |
| SE4     | 41.58         | 141.35         |
| SE5     | 41.64         | 141.32         |
| SE6     | 41.68         | 141.30         |
| SE7     | 41.74         | 141.26         |
| SE8     | 41.79         | 141.25         |
| SE9     | 41.83         | 141.22         |
Table 2. Implementation record of hydrographic observations along the Shiriya–Esan (SE) line.

The character of U, O, and W represent the observations conducted by T/S *Ushio-Maru*, T/S *Oshoro-Maru*, and R/V *Wakataka-Maru*, respectively.

| Year | Winter | Spring | Summer | Autumn |
|------|--------|--------|--------|--------|
|      | Jan.   | Feb.   | Mar.   | Apr.   | May    | Jun.   | Jul.   | Aug.   | Sep.   | Oct. | Nov. | Dec. |
| 2009 | U      | U      |        |        |        |        |        |        |        |      |      |      |
| 2010 | U      | U      |        |        |        |        |        |        |        |      |      | U    |
| 2011 | O      | U      |        |        |        |        |        |        |        |      |      | U    |
| 2012 | U      | U      |        |        |        |        |        |        |        |      |      | U    |
| 2013 | U      |        |        |        |        |        |        |        |        |      |      | U    |
| 2014 | U      | U      |        |        |        |        |        |        |        |      |      | U    |
| 2015 | U      | U      |        |        |        |        |        |        |        |      |      | U    |
| 2016 | U      | U      |        |        |        |        |        |        |        |      |      | U    |
| 2017 | U      |        |        |        |        |        |        |        |        |      |      |      |
| 2018 | U      |        |        |        |        |        |        |        |        |      |      | W    |
| 2019 | U      |        |        |        |        |        |        |        |        |      |      | U    |
Figures

Figure 1

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Supplementary Files

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- SuppFig1.jpg