Synergistic Effects of Midlatitude Atmospheric Upstream Disturbances and Oceanic Subtropical Front Intensity Variability on Western Pacific Jet Stream in Winter

Fei Fei Chen, Qiuyu Chen, Haibo Hu, Jiabei Fang, and Haokun Bai

Key Points:
- Strength and location of WPJS are determined by configuration of upstream disturbances and subtropical front intensity variability.
- Strong upstream disturbances anchor WPJS at about 25–45°N by obtaining baroclinic energy over this region.
- With weakened upstream disturbances, intensified subtropical front corresponds to southward movement of upward baroclinic energy and WPJS.

Abstract
In winter, the western Pacific jet stream (WPJS) is located at the downstream of East Asian subtropical jet (EASJ) and East Asian polar front jet (EAPF). It can be affected by sea surface temperature (SST) and has great impact on precipitation along the North American coast. Synergistic effects of midlatitude atmospheric upstream disturbances and oceanic subtropical front intensity variability on WPJS in winter are investigated in both observation and model experiments. It reveals that the strength and location of WPJS are determined by the configuration of atmospheric upstream disturbances and subtropical frontal zone (STFZ) intensity variability that exist concurrently. During the strong atmospheric upstream disturbances years, the WPJS is anchored at about 25–45°N, and its strength is enhanced (abated) by the intensified (weakened) STFZ. However, when the atmospheric upstream disturbances are weak, the intensity of STFZ changes the location of WPJS. An intensified (weakened) STFZ makes WPJS move southward (northward). Further studies show that strong atmospheric disturbances induce a wide and deep atmospheric baroclinicity anomaly zone, which makes the lower atmosphere obtain baroclinic energy from STFZ. Accordingly, the WPJS is anchored just the north of STFZ. At this time, the intensified (weakened) STFZ just increases (decreases) the upward baroclinic energy at the same position and thus the WPJS. However, with the weakened atmospheric upstream disturbances, the intensified (weakened) STFZ corresponds to an overhead narrow atmospheric baroclinicity anomaly zone, thus leading to the southward (northward) movements of upward baroclinic energy and the WPJS.

Plain Language Summary
The synergistic effect of atmospheric upstream disturbances and oceanic subtropical front intensity variability on western Pacific jet stream (WPJS) is not yet clear and is the key issue of this manuscript. Reanalysis data and model experiments are used to settle this question. The results suggest that the strength and location of WPJS are determined by the configuration of atmospheric upstream disturbances and subtropical frontal zone (STFZ) intensity variability. Strong atmospheric upstream disturbances anchor the WPJS at about 25–45°N by obtaining baroclinic energy from STFZ, while the strength of WPJS is affected by STFZ intensity. When the atmospheric upstream disturbances are weakened, the enhancement of STFZ makes both upward baroclinic energy and WPJS move southward. This study may be beneficial for precipitation prediction of North America.

1. Introduction
Oceanic fronts associated with sea surface temperature (SST) anomalies are considered to be key regions for the midlatitude ocean affecting the overlying atmosphere. Lots of studies have identified that oceanic fronts have a remarkable impact on the regional climate and weather system (G. Q. Liu et al., 2016; Small et al., 2008; Tokinaga et al., 2006; Vecchi et al., 2004; Wallace et al., 1990). For the basin scale, there is a significant negative correlation between SST and surface wind (Namias & Cayan, 1981). Negative (positive) SST anomalies often occur where the surface wind speed increases (weakens) (Wallace et al., 1990). However, a completely different SST-wind relationship was found over the small-scale SST front regions, which represents the forcing from the ocean to the atmosphere in the midlatitude (H. M. Xu et al., 2011; J. W. Liu et al., 2013; M. M. Xu & Xu, 2015; Tokinaga et al., 2006). A homomorphic relationship between SST and surface wind was observed in the Antarctic Circumpolar Current region (Annis & White, 2003), the Somali...
The two jets are strongest in winter, lying zonally along the northern and southern Panetta, 1993). The EASJ is an important component of the global subtropical jet, which is driven by the baroclinic zone, which is mainly formed by the eddy momentum
year, the East Asian subtropical jet (EASJ) and East Asian polar front jet (EAPJ). The EAPJ is located in the
In general, there are two branches of westerly jet streams in the upper troposphere over East Asia around the
year, the East Asian subtropical jet (EASJ) and East Asian polar front jet (EAPJ). The EAPJ is located in the
baroclinic zone, which is mainly formed by the eddy momentum flux convergence (Lee, 1997; Panetta, 1993). The EASJ is an important component of the global subtropical jet, which is driven by the
angular momentum transport along the poleward shift of the Hadley cell (Held & Hou, 1980; Hou, 1998).
The two jets are strongest in winter, lying zonally along the northern and southern flanks of the Tibetan Plateau (TP), respectively (D. Q. Huang, Zhu, et al., 2014; Luo & Zhang, 2015; Xiao & Zhang, 2012). EASJ
locates at about 20–35°N, south of the TP, and extends northeast to the North Pacific, while EAPJ locates
at about 40–60°N, north of the TP. In cool seasons, EASJ and EAPJ converge over the East Asian coastal
waters, and a jet generates and extends eastward into the oceanic region (Koch et al., 2006; Newton, 2004), which is known as WPJS. The shift of EASJ and EAPJ can reflect the interactions among different circulation systems (Schiemann et al., 2009; Y. C. Zhang & Huang, 2011) and therefore impact the local weather and climate as well as downstream regions (Guan et al., 2019; Y. Z. Zhang, Yan, et al., 2019; Zhou & Yu, 2005; Zhou & Zou, 2010; Zhu & Li, 2016). Branstator (2002) proved that the winter Asian jet could provide a waveguide for stationary Rossby waves and thus enhance the amplitude of low-frequency disturbance. Wave energy transfers to the downstream along with waveguide as well (Hoskins & Ambrizzi, 1993; Hu et al., 2018). Ren et al. (2009) found that two branches of the synoptic-scale transient eddy activity and low-level baroclinicity exist over the East Asian landmass, accompanied by the two-jet state of the EASJ and EAPJ in cool seasons. Synoptic-scale transient eddy activity accompanied by EASJ and EAPJ can be considered as the source of atmospheric upstream disturbances. Orlanski (2005) stressed the point that much of the storm track variability over the North Pacific can be understood as forced by variations in the upstream seeding of the storm track, which confirms the impact of atmospheric upstream disturbances on the midlatitude atmosphere over the North Pacific.

Overall, it is suggested that the interannual variability of the atmospheric upstream disturbances or STFZ intensity variability may influence the WPJS. But what are the synergistic effects of the atmospheric

SST anomalies in the oceanic front not only influence the surface wind speed but also play an important role in the formation and development of transient eddy in the storm tracks (Ma et al., 2017; Parfitt et al., 2016; Yao et al., 2016). The sensible heat flux (SHF) difference on both sides of the front can effectively maintain the near-surface baroclinicity and anchor the storm tracks in the midlatitude atmosphere (H. Nakamura et al., 2008). Taguchi and Nakamura (2009) and Taguchi et al. (2012) verified the significance of subarctic frontal zone (SAFZ) to the maintenance of storm tracks through oceanic baroclinic adjustment and pointed out the importance of the seasonality for the influence of the SAFZ on the atmosphere. The persistent advective effects of the monsoonal wind lead to weaker influences of SAFZ on the atmosphere in winter than in spring. In addition, the significance of SAFZ for wintertime western Pacific jet stream (WPJS) strength variability is demonstrated using observation data and global atmospheric numerical simulations (Ren et al., 2008). C. Zhang, Liu, et al. (2019) pointed out the variability of mesoscale SST in SAFZ impacts the turbulent heat fluxes out of the ocean, changing the near-surface baroclinicity with the large-scale zonal wind in the cold season. Albeit the subtropical frontal zone (STFZ) is weaker than SAFZ, the variability of STFZ can induce a significant atmospheric response (Kobashi et al., 2008). Wang et al. (2016, 2017) and L. Y. Zhang et al. (2017) investigated the relationship between the variability of STFZ intensity and WPJS in winter and spring, respectively. Results in observation suggest that the intensified STFZ leads to the reinforcement of WPJS in both winter and spring. Wang et al. (2019) pointed out that both the atmospheric upstream disturbances and STFZ intensity variability can influence wintertime WPJS strength in model experiments, and the WPJS intensity has a linear relationship with the STFZ intensity. Chen et al. (2019) fixed the upstream disturbances and found that in winter, atmospheric baroclinicity gets stronger rapidly through upward SHF after the STFZ enhancement. The strengthened storm tracks arouse more and stronger Rossby wave breaking (RWB) events in the upper troposphere, resulting in more transient perturbation kinetic energy transported to the WPJS, accelerating the westerly finally. However, as Wang et al. (2019) showed, the response of WPJS to STFZ is accompanied with upstream disturbances, which is not considered in the preliminary work of Chen et al. (2019).

In general, there are two branches of westerly jet streams in the upper troposphere over East Asia around the year, the East Asian subtropical jet (EASJ) and East Asian polar front jet (EAPJ). The EAPJ is located in the baroclinic zone, which is mainly formed by the eddy momentum flux convergence (Lee, 1997; Panetta, 1993). The EASJ is an important component of the global subtropical jet, which is driven by the angular momentum transport along the poleward shift of the Hadley cell (Held & Hou, 1980; Hou, 1998). The two jets are strongest in winter, lying zonally along the northern and southern flanks of the Tibetan Plateau (TP), respectively (D. Q. Huang, Zhu, et al., 2014; Luo & Zhang, 2015; Xiao & Zhang, 2012). EASJ locates at about 20–35°N, south of the TP, and extends northeast to the North Pacific, while EAPJ locates at about 40–60°N, north of the TP. In cool seasons, EASJ and EAPJ converge over the East Asian coastal waters, and a jet generates and extends eastward into the oceanic region (Koch et al., 2006; Newton, 2004), which is known as WPJS. The shift of EASJ and EAPJ can reflect the interactions among different circulation systems (Schiemann et al., 2009; Y. C. Zhang & Huang, 2011) and therefore impact the local weather and climate as well as downstream regions (Guan et al., 2019; Y. Z. Zhang, Yan, et al., 2019; Zhou & Yu, 2005; Zhou & Zou, 2010; Zhu & Li, 2016). Branstator (2002) proved that the winter Asian jet could provide a waveguide for stationary Rossby waves and thus enhance the amplitude of low-frequency disturbance. Wave energy transfers to the downstream along with waveguide as well (Hoskins & Ambrizzi, 1993; Hu et al., 2018). Ren et al. (2009) found that two branches of the synoptic-scale transient eddy activity and low-level baroclinicity exist over the East Asian landmass, accompanied by the two-jet state of the EASJ and EAPJ in cool seasons. Synoptic-scale transient eddy activity accompanied by EASJ and EAPJ can be considered as the source of atmospheric upstream disturbances. Orlanski (2005) stressed the point that much of the storm track variability over the North Pacific can be understood as forced by variations in the upstream seeding of the storm track, which confirms the impact of atmospheric upstream disturbances on the midlatitude atmosphere over the North Pacific.
upstream disturbances and STFZ intensity variabilities on the WPJS in winter considering the concurrent variations at the interannual time scale? Whether the strength or location of WPJS will be changed? The precipitation along the North American coast is quite different with the WPJS changes in strength or location (Ellis & Barton, 2012); reasons for the changes of strength or location of WPJS are still unclear. To settle the questions above, observation data and atmospheric regional model were used to analyze the synergistic effects of atmospheric upstream disturbances and STFZ intensity variability on WPJS. The rest of the paper is organized as follows. Data and method are described in section 2. We find out the synergistic effects in observation and model experiments and analyze the mechanism in section 3 and section 4. The summary and discussion are presented in section 5.

2. Data and Method

2.1. Observation Data
The SST data set used in this paper is the Climate Forecast System Reanalysis (CFSR) 6-hourly SST data, provided by National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR), with a spatial resolution of 0.5° × 0.5°, for winter (December–February) from 1979 to 2015. Besides, NCEP/CFSR 6-hourly atmospheric data set with a spatial resolution of 0.5° × 0.5° and 37 vertical pressure levels was used to describe the change of midlatitude atmosphere during the wintertime of 1979–2015.

2.2. Model Experiments
We chose Weather Research and Forecasting Model Version 3.5 (WRF 3.5) as the atmospheric regional model, which is developed by NCEP/NCAR and other institutions. The WRF model is one of the most advanced regional meteorological models and is widely recognized for studies of the medium and small-scale air-sea interaction (H. M. Xu et al., 2011; Kilpatrick et al., 2016; O’Neill et al., 2010; Skamarock et al., 2008). Some vital physical parameterization schemes used in the model include the Thompson scheme (Thompson et al., 2004) for microphysical parameterization, the Betts-Miller-Janjic scheme (Janjić, 1994, 2000) for cumulus parameterization, and the Yonsei University (YSU) atmospheric boundary layer parameterization scheme (Hong et al., 2006). The high-resolution CFSR data were used as the underlying and lateral boundary conditions for our simulations. The model domain was 0.5°–59.5°N, 127°E to 122°W, with a horizontal resolution of 50 km and 28 sigma levels in vertical. The model output data were obtained every 6 hr.

To test the credibility of the differences between different experiment results, we use the T test and assume that there are m samples with one experiment and n simples with the other one. The mean value of the meteorological element X of the two sample groups we calculated are \( \bar{X}_m \) and \( \bar{X}_n \). With the null hypothesis that the overall averages of the two samples have no significant differences, the statistics t follows t distribution with \( (m+n-2) \) degrees of freedom. The \( s^2 \) in it means the unbiased estimator of the variance of the difference. \( \bar{X}_m \) and \( \bar{X}_n \) mean the average of each experiment results respectively (Chen et al., 2019; Chervin & Schenider, 1976; Trenberth, 1984).

\[
\begin{align*}
    t &= \frac{\bar{X}_m - \bar{X}_n}{\sqrt{\frac{1}{m} + \frac{1}{n}}} \\
    s^2 &= \frac{\sum_{i=1}^{m} (X_i - \bar{X}_m)^2 + \sum_{i=1}^{n} (X_i - \bar{X}_n)^2}{m + n - 2}
\end{align*}
\]

For given significant level \( (\alpha) \) and degrees of freedom \( (m+n-2) \), we can find the corresponding critical value \( t_{\alpha} \). If \( t > t_{\alpha} \), we reject the null hypothesis, and we think that the differences are significant.

3. Observed Synergistic Effects of Midlatitude Atmospheric Upstream Disturbances and STFZ Intensity Variability on WPJS

3.1. Atmospheric Anomalies Associated With the Synergistic Effects
The number of climatological jet cores at 250 hPa is shown in Figure 1a. EASJ and EAPJ converge at about 128°E, and the jet extends eastward into the oceanic region. The jet over the North Pacific region is often
called WPJS, which shows its maximum speed in vicinity of Japan. Wang et al. (2019) pointed out that both the atmospheric upstream disturbances and STFZ intensity variability can influence the WPJS, so we first define their intensity indices. Figure 1b shows the climatological mean and variance of eddy kinetic energy (EKE) at 127–129°E, where EASJ and EAPJ converge. We define the averaged EKE within the purple frame (20–55°N, 100–700 hPa) as the intensity index of atmospheric upstream disturbances, where both the climatological mean and the variance of EKE are strong. A year with its standardized intensity index greater than 1.0 (less than −1.0) is defined as a strong (weak) atmospheric upstream disturbances year. The selected strong atmospheric upstream disturbances years include 1986, 1991, 1994, 1997, 2001, 2003, 2004, 2009, 2010, and 2015, while weak atmospheric upstream disturbances years are 1980, 1993, 1995, 2000, 2007, 2011, and 2012 (Figure 2a).

STFZ distributed in a meridionally narrow and zonally elongated manner over the North Pacific basin (140°E to 140°W) at about 24–32°N (Figure 1c). Similar to Wang et al. (2016), we define the averaged SST gradient within the STFZ (140°E to 140°W, 24–32°N) as the intensity index of STFZ. A standardized intensity index greater than 1.0 (less than −1.0) represents a strong (weak) STFZ year. Strong STFZ years include 1980, 1986, 1995, 2002, 2005, 2009, and 2014, while weak STFZ years are 1990, 1992, 1994, 1999, 2008, and 2011 (Figure 2b).

Combined with the intensity index of atmospheric upstream disturbances and STFZ, we define four types of events in observation: (1) strong upstream disturbances and strong STFZ (OBS_STR_WEST&STR_STFZ): the standardized upstream disturbances index greater than 1.0 and the standardized STFZ index greater than 0.0; (2) strong upstream disturbances and weak STFZ (OBS_STR_WEST&WEA_STFZ): the
standardized upstream disturbances index greater than 1.0 and the standardized STFZ index less than 0.0; (3) weak upstream disturbances and strong STFZ (OBS_WEASY&STR_STFZ): the standardized upstream disturbances index less than −1.0 and the standardized STFZ index greater than 0.0; and (4) weak upstream disturbances and weak STFZ (OBS_WEASY&WEA_STFZ): the standardized upstream disturbances index less than −1.0 and the standardized STFZ index less than 0.0. Specific years corresponding to the four types of events are shown in Table 1.

We use composite analysis to find out synergistic effects of atmospheric upstream disturbances and STFZ intensity variability on WPJS. Differences of geopotential height and zonal wind between OBS_STR_WEST&STR_STFZ and OBS_STR_WEST, and OBS_STR_WEST&WEA_STFZ and OBS_STR_WEST years are shown in Figure 3. When atmospheric upstream disturbances are strong, 250 hPa zonal wind anomalies appear downstream of the WPJS main body (Figures 3a and 3b). Intensified STFZ corresponds to equivalent barotropic strength changes of the WPJS at about 25–45°N, close to the large value region of climatological mean zonal wind, while weakened STFZ shows the opposite effects (Figures 3c–3f). Differences of geopotential height and zonal wind between OBS_WEASY&WSTFZ and OBS_WEASY, and OBS_WEASY&WSTFZ and OBS_WEASY years are shown in Figure 4. When atmospheric upstream disturbances are weakened, 250 hPa zonal wind field shows the southward (northward) shift of the WPJS (Figures 4a and 4b). Intensified STFZ corresponds to stronger WPJS at about 20–35°N, more southerly of the large value region of climatological mean zonal wind, while weakened STFZ shows the opposite effects (Figures 4c–4f). Comparing the results of Figures 3 and 4, we can conclude that strong atmospheric upstream disturbances can anchor the WPJS at about 25–45°N, corresponding to the large value region of climatological WPJS. The strength of WPJS is affected by STFZ intensity, while intensified (weakened) STFZ corresponds to stronger (weaker) WPJS. However, when the atmospheric upstream disturbances are weakened, the latitudinal position of WPJS is changed by STFZ. The enhancement of STFZ makes WPJS move southward to about 20–35°N, corresponding to the location of STFZ, more southerly than the large value region of climatological WPJS, and vice versa.

Table 1

| Events                                                                 | Years                          |
|-----------------------------------------------------------------------|-------------------------------|
| OBS_STR_WEST&STR_STFZ (strong atmospheric upstream disturbances and strong STFZ in observation) | 1986, 1991, 1997, 2003, 2009, and 2015 |
| OBS_STR_WEST&WEA_STFZ (strong atmospheric upstream disturbances and weak STFZ in observation) | 1994, 2001, 2004, and 2010    |
| OBS_WEASY&WSTFZ (weak atmospheric upstream disturbances and strong STFZ in observation) | 1980, 1995, and 2000          |
| OBS_WEASY&WSTFZ (weak atmospheric upstream disturbances and weak STFZ in observation) | 1993, 2007, 2011, and 2012    |

Figure 2. The standardized intensity index of (a) atmospheric upstream disturbances and (b) STFZ from 1979 to 2015.
Figure 3. Differences of zonal wind (shadings, unit: m/s) at 250 hPa between (a) OBS_STR_WEST&STR_STFZ and OBS_STR_WSET and (b) OBS_STR_WEST&WEA_STFZ and OBS_STR_WEST years. The purple frame is the location of STFZ. (c and d) The same as (a) and (b) but for differences of zonally averaged zonal wind (shadings, unit: m/s) over 140°E to 140°W. (e and f) The same as (c) and (d) but for differences of $z^2$ (shadings, unit: m$^2$). The red triangle means the central latitude of STFZ. Contours are the climatological mean of each physical quantity in winter from 1979 to 2015. Areas marked by dots passed the 0.01 significance level, and the areas marked by slashes passed the 0.05 significance level.
Figure 4. Differences of zonal wind (shadings, unit: m/s) at 250 hPa between (a) OBS_WEA_WEST&STR_STFZ and OBS_WEA_WSET and (b) OBS_WEA_WEST&WEA_STFZ and OBS_WEA_WEST years. The purple frame is the location of STFZ. (c and d) The same as (a) and (b) but for differences of zonally averaged zonal wind (shadings, unit: m/s) over 140°E to 140°W. (e and f) The same as (c) and (d) but for differences of $z^2$ (shadings, unit: m$^2$). The red triangle means the central latitude of STFZ. Contours are the climatological mean of each physical quantity in winter from 1979 to 2015. Areas marked by dots passed the 0.01 significance level, and the areas marked by slashes passed the 0.05 significance level.
Variations of the jet stream can further affect the downstream precipitation (A. N. Huang, Zhou, et al., 2014; Ellis & Barton, 2012; Simpson et al., 2019). As shown in Figure 5, when the atmospheric upstream disturbances are strong, precipitation anomalies are observed over the midlatitude basin between 30°N and 50°N, in particular the northwestern coast of North America (Figures 5a and 5b). When the atmospheric upstream disturbances are weakened, the enhancement of STFZ corresponds to the southward movement of WPJS, and the positive precipitation anomalies move southward to 20°–45°N, and vice versa (Figures 5c and 5d). However, the influences of STFZ on precipitation are different with different upstream disturbances. When the atmospheric upstream disturbances are strong, strong (weak) STFZ only accelerates (decreases) the WPJS without changing the jet location, thus increasing (decreasing) the zonal moisture transportation and leads to anomalous precipitation at the exit region of the jet (Figures 5a and 5b). Nevertheless, when the atmospheric upstream disturbances are weak, the enhancement of STFZ corresponds to the southward movement of both WPJS and the positive precipitation anomalies, causing the decrease of the poleward transportation of moisture from low latitudes. As a result, the induced precipitation anomalies with weak upstream disturbances are more significant than that with strong disturbances, and vice versa (Figures 5c and 5d). It is suggested that midlatitude atmospheric upstream disturbances and STFZ can affect precipitation along the North American coast, which may be beneficial for precipitation prediction and disaster prevention in North America.

3.2. Mechanism Analysis of the Synergistic Effects

We first analyze the mechanism of synergistic effects of midlatitude atmospheric upstream disturbances and STFZ intensity variability on WPJS in observation. Near-surface baroclinicity is crucial to the baroclinic growth of synoptic disturbances through enhancing the vertical coupling between incoming upper-level eddies and potential vorticity anomalies induced thermally at the surface (Hoskins et al., 1985; H. Nakamura & Sampe, 2002). Therefore, we first examine the 850 hPa poleward eddy heat flux. When atmospheric upstream disturbances are strong, intensified STFZ corresponds to stronger poleward eddy heat flux at about 20°–45°N over the North Pacific basin, while weakened STFZ shows the opposite effects (Figures 6a and 6b). When atmospheric upstream disturbances are weakened, the poleward eddy heat flux differences locate along the STFZ, more southerly than the differences in strong upstream disturbances years.
Differences of zonally averaged poleward eddy heat flux over 140°E to 140°W are shown in Figure 7. When the atmospheric upstream disturbances are strong, intensified STFZ corresponds to stronger poleward eddy heat flux throughout the tropopause level at about 20–45°N (Figures 7a and 7b). When the atmospheric upstream disturbances are weakened, the poleward eddy heat flux differences locate along the STFZ at lower levels and extend to about 35°N at 250 hPa (Figures 7c and 7d).

The stronger vertical coupling of eddies and potential vorticity leads to enhanced baroclinic growth. The formula of the baroclinic eddy growth rate (Hoskins & Valdes, 1990; Nakamura & Yamane, 2010) is as follows:

$$\sigma_{Bl} = 0.31 \frac{g}{N} \left| \frac{\partial \theta}{\partial y} \right|$$  \hspace{1cm} (1)

where $N^2 = -g \alpha \frac{\partial \theta}{\partial p}$, $\alpha = \frac{T}{\frac{\partial T}{\partial p} \mid_{p=1000}}$, $g$ is the gravitational acceleration, $\theta$ is the potential temperature, $p$ means pressure, and $T_p \mid_{p=1000}$ represents the atmospheric temperature at 1,000 hPa. Differences of baroclinic eddy growth rate between different events are shown in Figure 8. When atmospheric upstream disturbances are strong, intensified STFZ corresponds to stronger atmospheric baroclinicity at about 20–45°N (Figure 8a). This means that the strong atmospheric upstream disturbances anchor the upward baroclinic energy at about 20–45°N. When atmospheric upstream disturbances are weakened, intensified STFZ corresponds to stronger baroclinicity at about 20–35°N at lower levels and then extending to upper troposphere (Figure 8c). Intensified (Weakened) atmospheric baroclinicity leads to more (less) active transient eddies, which can influence the intensity of WPJS.

Eliassen-Palm (E-P) flux is an indicator of eddy activity, and its horizontal divergence is related to the acceleration of WPJS (Hoskins et al., 1983; Trenberth, 1986). The formula of E-P flux is as follows:

Figure 6. Differences of $\frac{\partial \theta}{\partial y}$ (shadings, unit: K·m/s) at 850 hPa between (a) OBS_STR_WEST&STR_STFZ and OBS_STR_WEST, (b) OBS_STR_WEST&WEA_STFZ and OBS_STR_WEST, (c) OBS_WEAL_WEST&STR_STFZ and OBS_WEAL_WEST, and (d) OBS_WEAL_WEST&WEA_STFZ and OBS_WEAL_WEST years. Areas marked by dots passed the 0.01 significance level, and the areas marked by slashes passed the 0.05 significance level. The purple frame is the location of STFZ.
\[ E = \left( \frac{\vec{v'}^2 - \vec{u'}^2}{2}, -\vec{u'}\vec{v'} \int \frac{\vec{v}\partial\theta}{\partial\theta} \cos\phi \right) \cos\phi \] (2)

where \( u', v', \) and \( \theta' \) mean transient disturbance of zonal wind, meridional wind, and potential temperature, respectively, \( \phi \) means the latitude, and \( p \) means pressure. The symbol \( \cdashdot \) means the time average of relative variables. \( f \) is the Coriolis force and \( f = 2\Omega \sin\phi \), where \( \Omega \) is the Coriolis parameter, \( \Omega = 7.29 \times 10^{-5}/s \).

To eliminate the influence of SAFZ, we choose 180°–150°W to analyze the vertical E-P flux anomalies. The horizontal divergence of E-P flux is basically consistent with WPJS (Figure 9). Strong atmospheric upstream disturbances anchor the upward E-P flux at about 30°–45°N (Figures 9a and 9b). When the upstream disturbances are weakened, intensified STFZ corresponds to upward E-P flux at about 20°–35°N at lower levels, extending to upper troposphere (Figures 9c and 9d). Combining the results above, we conclude that strong atmospheric disturbances make the lower atmosphere obtain baroclinic energy from STFZ, thereby anchoring the WPJS to the north of STFZ. At this time, the intensified (weakened) STFZ influence the WPJS strength by increasing (decreasing) the upward baroclinic energy, but the latitudinal position of the upward baroclinic energy is anchored. When the atmospheric upstream disturbances are weakened, the enhancement of STFZ also corresponds to the southward movements of the WPJS and associated upward baroclinic energy, and vice versa.
4. Simulated WPJS Responses to the Synergistic Effects of Midlatitude Atmospheric Upstream Disturbances and STFZ Intensity Variability

We design four experiments to verify the synergistic effects of atmospheric upstream disturbances and STFZ intensity variability on WPJS (Table 2). (1) EXP_STR_WEST_CLI: The atmospheric boundary condition is the NCEP/CFSR 6-hourly winter data in 1986. The ocean boundary condition is the reconstructed idealized SST (Figure 10a), that is, extending the zonally averaged SST over 140°E to 140°W in winter from 2000 to 2009 to the whole Pacific basin to obtain zonally uniform SST with only meridional gradients.

(2) EXP_STR_STFZ_STR: The atmospheric boundary condition is the same as the EXP_STR_WEST_CLI experiment. To eliminate the influence of model initial disturbance, we use the ocean boundary condition of EXP_STR_WEST_CLI experiment as a basis and add the specific SST anomalies (Figure 10b) into the STFZ at 6:00 a.m. on 21 December.

(3) EXP_WEA_WEST_CLI: The atmospheric boundary condition is the NCEP/CFSR 6-hourly winter data in 2011. The ocean boundary condition is the same as the EXP_STR_WEST_CLI experiment.

(4) EXP_WEA_WEST_STR: The atmospheric boundary condition is the same as the EXP_WEA_WEST_CLI experiment. The ocean boundary condition is the same as the EXP_STR_WEST_STR experiment. The integral time of the four experiments are from 0:00 a.m. on 1 December to 6:00 p.m. on 28 February next year. To avoid any influence of the original state on the model initial state in the STFZ, the observation of EXP_STR_WEST_CLI and EXP_STR_STFZ_STR is subtracted from OBS_STR_WEST and OBS_STR_STFZ, respectively.

Figure 8. Differences of zonally averaged baroclinic eddy growth rate (shadings, unit: $10^{-6} \text{ m/s}^3$) and zonal wind (contours, unit: m/s) over 140°E to 140°W between (a) OBS_STR_WEST&STR_STFZ and OBS_STR_WEST, (b) OBS_STR_WEST&WEA_STFZ and OBS_STR_WEST, (c) OBS_WEA_WEST&STR_STFZ and OBS_WEA_WEST, and (d) OBS_WEA_WEST&WEA_STFZ and OBS_WEA_WEST years. The red triangle means the central latitude of STFZ. Areas marked by dots passed the 0.01 significance level, and the areas marked by slashes passed the 0.05 significance level.
Simulation results, we carried out different initialization experiments every 24 hr from 0:00 a.m. on 1 December to 0:00 a.m. on 5 December in each simulation (five times each group with different initialization). All results were based on the ensemble means of the five runs in each group.

**Figure 9.** Differences of zonally averaged E-P flux (vectors, unit: m²/s²), zonal wind (contours, unit: m/s), and horizontal divergence of E-P flux (shadings, unit: m²/s²) over 180°–150°W between (a) OBS_STR_WEST&STR_STFZ and OBS_STR_WEST, (b) OBS_STR_WEST&WEA_STFZ and OBS_STR_WEST, (c) OBS_WEA_WEST&STR_STFZ and OBS_WEA_WEST, and (d) OBS_WEA_WEST&WEA_STFZ and OBS_WEA_WEST years. The red triangle means the central latitude of STFZ.

**Table 2**  
Introduction to Experiment Design

| Experiment name                                           | Specific experiment design                                      |
|------------------------------------------------------------|-----------------------------------------------------------------|
| EXP_STR_WEST_CTL (strong atmospheric western boundary and reconstructed idealized STFZ in model experiment) | Reconstructed idealized SST (Figure 10a) + 1986 NCEP/CFSR 6-hourly winter atmospheric lateral boundary |
| EXP_STR_WEST_STR (strong atmospheric western boundary and reconstructed enhanced STFZ in model experiment) | Reconstructed enhanced SST (Figure 10c) + 1986 NCEP/CFSR 6-hourly winter atmospheric lateral boundary |
| EXP_WEA_WEST_CTL (weak atmospheric western boundary and reconstructed idealized STFZ in model experiment) | Reconstructed idealized SST (Figure 10a) + 2011 NCEP/CFSR 6-hourly winter atmospheric lateral boundary |
| EXP_WEA_WEST_STR (weak atmospheric western boundary and reconstructed enhanced STFZ in model experiment) | Reconstructed enhanced SST (Figure 10c) + 2011 NCEP/CFSR 6-hourly winter atmospheric lateral boundary |
Similar to the above section, differences of poleward eddy heat flux and E-P flux between EXP_STR_WEST_STR and EXP_STR_WEST_CLI, and EXP_WEA_WEST_STR and EXP_WEA_WEST_CLI experiments are shown in Figure 11. When atmospheric upstream disturbances are strong, intensified STFZ corresponds to stronger poleward eddy heat flux throughout the tropopause level at about 20–45°N over the North Pacific basin (Figures 11a and 11c). When atmospheric upstream disturbances are weakened, the poleward eddy heat flux differences locate along the STFZ, more southerly than the differences in strong upstream disturbances years (Figure 11b). Besides, strong atmospheric upstream disturbances anchor the upward E-P flux at about 30–45°N (Figure 11e). When the upstream disturbances are weakened, intensified STFZ corresponds to upward E-P flux at about 20–35°N at lower levels, extending to upper troposphere (Figure 11f).

Differences of zonal wind between EXP_STR_WEST_STR and EXP_STR_WEST_CLI, and EXP_WEA_WEST_STR and EXP_WEA_WEST_CLI experiments are shown in Figure 12. When atmospheric upstream disturbances are strong, the WPJS gets stronger at about 25–45°N, corresponding to the large value region of climatological zonal wind (Figures 12a and 12c). When atmospheric upstream disturbances are weakened, intensified STFZ corresponds to stronger WPJS at about 20–35°N, corresponding to the location of STFZ, more southerly than the large value region of climatological zonal wind (Figures 12b and 12d).
Figure 11. Differences of $\nabla T'$ (shadings, unit: K·m/s) at 850 hPa between (a) EXP_STR_WEST_STR and EXP_STR_WSET_CLI and (b) EXP_WEA_WEST_STR and EXP_WEA_WEST_CLI experiments. The purple frame is the location of STFZ. (c and d) The same as (a) and (b) but for differences of zonally averaged $\nabla T'$ (shadings, unit: K·m/s) over 140°E to 140°W. (e and f) The same as (c) and (d) but for differences of E-P flux (vectors, unit: m²/s²) and horizontal divergence of E-P flux (shadings, unit: m/s²) over 140°E to 140°W. The red triangle means the central latitude of STFZ. Areas marked by dots passed the 0.01 significance level, and the areas marked by slashes passed the 0.05 significance level.
Results of the experiments are somewhat different from observation. The 250 hPa zonal wind anomaly appears at the WPJS main body in simulation, more westerly than observation. When the atmospheric upstream disturbances are weakened, the WPJS anomalies are more southerly than observation which may be influenced by lots of factors (Jiang et al., 2019). Lots of previous studies have mentioned that large-scale SST anomalies (H. Nakamura et al., 1997; Wang et al., 2019), SAFZ and STFZ intensity (Chen et al., 2020; Hoskins & Valdes, 1990; L. Y. Zhang et al., 2017; Ren et al., 2008; Wang et al., 2016), ocean eddies (Wang et al., 2020; Wen et al., 2020), atmospheric upstream disturbances (Chu et al., 2020; Zhang et al., 2012), and land surface processes (Xue et al., 2004) all have great influences on WPJS. However, in this study, we only focus on the effects of STFZ intensity accompany with the atmospheric upstream disturbances on the WPJS. The ideal SST used in our model simulations only keeps STFZ intensity the same as observation, but with different initial SST field (which has been zonal averaged). In addition, other forcing terms in observation were not considered in our simulations either. Therefore, the simulation results are not exactly the same but very close to the observation. Overall, the simulation results support what we conclude in observation. Strong atmospheric disturbances make the lower atmosphere obtaining baroclinic energy from STFZ, thereby anchoring the WPJS at about 25–45°N. At this time, the intensified (weakened) STFZ influences the WPJS strength by increasing (decreasing) the upward baroclinic energy, but the latitudinal position of the upward baroclinic energy is anchored. When the atmospheric upstream disturbances are weakened, the enhancement of STFZ makes upward baroclinic energy and WPJS move southward to about 20–40°N, corresponding to the location of STFZ, and vice versa.

Figure 12. Differences of zonal wind (shadings, unit: m/s) at 250 hPa between (a) EXP_STR_WEST_STR and EXP_STR_WSET_CLI and (b) EXP_WEA_WEST_STR and EXP_WEA_WEST_CLI experiments. (c and d) The same as (a) and (b) but for differences of zonally averaged zonal wind (shadings, unit: m/s) over 140°E to 140°W. The red triangle means the central latitude of STFZ. Contours are the climatological mean of each physical quantity in winter. Areas marked by dots passed the 0.01 significance level, and the areas marked by slashes passed the 0.05 significance level.
5. Summary and Discussion

In this study, we focused on the synergistic effects of midlatitude atmospheric upstream disturbances and STFZ intensity variability on WPJS in winter. First, combined with the intensity index of atmospheric upstream disturbances and STFZ, we define four types of events: OBS_STR_WEST&STR_STFZ, OBS_STR_WEST&WEA_STFZ, OBS_WEA_WEST&STR_STFZ, and OBS_WEA_WEST&WEA_STFZ years. Then, we used composite analysis to find out synergistic effects of atmospheric upstream disturbances and STFZ intensity variability on WPJS in observation. We found that strong atmospheric upstream disturbances can anchor the WPJS at about 25–45°N, corresponding to the large value region of climatological WPJS. The strength of WPJS is affected by STFZ intensity, while intensified (weakened) STFZ corresponds to stronger (weaker) WPJS. However, when the atmospheric upstream disturbances are weakened, the latitudinal position of WPJS is changed by STFZ. The enhancement of STFZ makes WPJS move southward to about 20–35°N, more southerly than the large value region of climatological WPJS, and vice versa. After that, we analyzed the mechanism of synergistic effects of midlatitude atmospheric upstream disturbances and STFZ intensity variability on WPJS and find that the change of STFZ intensity can affect the upper WPJS with both strong upstream disturbances and weak upstream disturbances, but in somewhat different ways. Strong atmospheric disturbances make the lower atmosphere obtain baroclinic energy from STFZ, thereby anchoring the WPJS to the north of STFZ. At this time, the...

Figure 13. Schematic diagrams of the synergistic effects in (a) OBS_STR_WEST and (b) OBS_WEA_WEST years. SST (shadings, unit: K), SST gradient (curves, unit: °C/100 km), climatological mean zonal wind (vectors, unit: m/s), zonal wind anomaly (shadings, unit: m/s) at 250 hPa, and zonally averaged E-P flux (vectors, unit: m²/s²) over 180–150°W are shown.

10.1029/2020JD032788

Journal of Geophysical Research: Atmospheres
intensified (weakened) STFZ influence the WPJS strength by increasing (decreasing) the upward baroclinic energy, but the latitudinal position of the upward baroclinic energy is anchored. When the atmospheric upstream disturbances are weakened, the enhancement of STFZ corresponds to the southward movement of upward baroclinic energy and the WPJS, and vice versa. To verify the phenomenon and mechanism in observation, we design four experiments: EXP_STR_WEST_CLI, EXP_STR_WEST_STR, EXP_WEAST_CLI, and EXP_WEAST_STR. Results of the experiments confirm what we conclude in observation. The specific processes are shown in Figure 13.

Different from the previous study, we focus on the synergistic effects of midlatitude atmospheric upstream disturbances and STFZ intensity variability. Chen et al. (2019) proved that enhanced STFZ can lead to a stronger upper zonal wind, which is consistent with our results, but ignores the changes in the location of WPJS. In this study, we found that strong atmospheric upstream disturbances anchor the WPJS at about 25°–45°N. And when the atmospheric upstream disturbances are weakened, the enhancement of STFZ corresponds to the southward movement of WPJS. However, limited by the length of the CFSR data set, samples of four types of events are not enough; a long-term high-resolution air-sea coupled data set is needed for future research. Although STFZ intensity variability has been proved to have significant effect on WPJS on the interannual time scale, SAFZ also has great impact on WPJS considering its larger temperature gradient (H. Nakamura et al., 2008; Taguchi & Nakamura, 2009). Differences of how the two oceanic fronts influence the WPJS will be carried out in future research.

**Data Availability Statement**

The NCEP/CFSR reanalysis data used in this study is obtained from the website (https://rda.ucar.edu/datasets/ds093.0/).

**References**

Annis, J. L., & White, W. B. (2003). Coupling of extratropical mesoscale eddies in the ocean to westerly winds in the atmospheric boundary layer. *Journal of Physical Oceanography*, 33(5), 1095–1107. https://doi.org/10.1175/1520-0469(2003)033<1095:CEMTEO>2.0.CO;2

Bai, H. K., Hu, H. B., Perrie, W., & Zhang, N. (2020). On the characteristics and climate effects of HV-WCP events over the Kuroshio SST front during wintertime. *Climate Dynamics*. https://doi.org/10.1007/s00382-020-05373-5

Bai, H. K., Hu, H. B., Yang, X. Q., Ren, X. J., Xu, H. M., & Liu, G. Q. (2019). Modeled MABL responses to the winter Kuroshio SST front in the East China Sea and Yellow Sea. *Journal of Geophysical Research: Atmospheres*, 124, 6069–6092. https://doi.org/10.1029/2018JD032970

Branstator, G. (2002). Circumglobal teleconnections, the jet stream waveguide, and the North Atlantic Oscillation. *Journal of Climate*, 15(4), 1893–1910. https://doi.org/10.1175/1520-0442(2002)015<1893:CMTJST>2.0.CO;2

Chen, F. F., Hu, H. B., & Bai, H. K. (2020). Subseasonal coupling between subsurface subtropical front and overlying atmosphere in North Pacific winter. *Dynamics of Atmospheres and Oceans*, 90(11145), 101145. https://doi.org/10.1016/j.dynatmoce.2020.101145

Chen, Q. Y., Hu, H. B., Ren, X. J., & Yang, X. Q. (2019). Numerical simulation of midlatitude upper-level zonal wind response to the change of North Pacific subtropical front strength. *Journal of Geophysical Research: Atmospheres*, 124, 4891–4912. https://doi.org/10.1029/2018JD029589

Chervin, R. M., & Schenider, S. H. (1976). On determining the statistical significance of climate experiments with general circulation models. *Journal of the Atmospheric Sciences*, 33(3), 405–412. https://doi.org/10.1175/1520-0469(1976)033<0405:COEMEI>2.0.CO;2

Chu, C. J., Hu, H. B., Yang, X. Q., & Yang, D. J. (2020). Midlatitude atmospheric transient eddy feedbacks influenced ENSO-associated wintertime Pacific teleconnection patterns in two PDO phases. *Climate Dynamics*, 54(1-3), 2577–2595. https://doi.org/10.1007/s00382-020-05134-4

Ellis, A. W., & Barton, N. P. (2012). Characterizing the north pacific jet stream for understanding historical variability in western United States winter precipitation. *Physical Geography*, 33(2), 105–128. https://doi.org/10.2747/0272-3646.33.2.105

Guo, W. N., Hu, H. B., Ren, X. J., & Yang, X. Q. (2019). Subseasonal zonal variability of the western pacific subtropical high in summer: Climate impacts and underlying mechanisms. *Climate Dynamics*, 51, 3325–3344. https://doi.org/10.1007/s00382-019-04705-4

Held, I. M., & Hou, A. Y. (1986). Nonlinear axially symmetric circulations in a nearly inviscid atmosphere. *Journal of the Atmospheric Sciences*, 3(7), 515–533. https://doi.org/10.1175/1520-0469(1980)037<0515:NASCIA>2.0.CO;2

Hirata, H., Kawamura, R., Kato, M., & Shinoda, T. (2016). Response of rapidly developing extratropical cyclones to sea surface temperature variations over the western Kuroshio-Oyashio confluence region: Extratropical cyclone and SST variations. *Journal of Geophysical Research: Atmospheres*, 121, 3842–3858. https://doi.org/10.1002/2013JD024391

Hong, S., Noh, Y., & Dudhia, J. (2006). A new vertical diffusion package with an explicit treatment of entrainment processes. *Monthly Weather Review*, 134(9), 2318–2341. https://doi.org/10.1175/MWR3199.1

Hoskins, B. J., & Ambrizzi, T. (1993). Rossby wave propagation on a realistic longitudinally varying flow. *Journal of the Atmospheric Sciences*, 50(12), 1661–1671. https://doi.org/10.1175/1520-0469(1993)050<1661:RWPOAR>2.0.CO;2

Hoskins, B. J., James, I. N., & White, G. H. (1983). The shape, propagation and mean-flow interaction of large-scale weather systems. *Journal of the Atmospheric Sciences*, 40(7), 1595–1612. https://doi.org/10.1175/1520-0469(1983)040<1595:TSPIAM>2.0.CO;2

Hoskins, B. J., McIntyre, M. E., & Robertson, A. W. (1985). On the use and significance of isentropic potential vorticity maps. *Quarterly Journal of the Royal Meteorological Society*, 111(470), 877–946. https://doi.org/10.1002/qj.49711147002

Hoskins, B. J., & Valdes, P. J. (1999). On the existence of storm-tracks. *Journal of the Atmospheric Sciences*, 47(15), 1854–1864. https://doi.org/10.1175/1520-0469(1999)047<1854:OETOEST>2.0.CO;2
Taguchi, B., Nakamura, H., Nonaka, M., Komori, N., Kewano-Yoshida, A., Takaya, K., & Goto, A. (2012). Seasonal evolutions of atmospheric response to decadal SST anomalies in the North Pacific subarctic front zone: Observations and a coupled model simulation. *Journal of Climate*, 25(1), 111–139. https://doi.org/10.1175/JCLI-D-11-00046.1

Thompson, G., Rasmussen, R., & Manning, K. W. (2004). Explicit forecasts of winter precipitation using an improved bulk microphysics scheme. Part I: Description and sensitivity analysis. *Monthly Weather Review*, 132(2), 519–542. https://doi.org/10.1175/1520-0493(2004)132<0519:EFPWSI>2.0.CO;2

Tokinaga, H., Tanimoto, Y., Nonaka, M., Taguchi, B., Fukumachi, T., Xie, S. P., & Nakamura, H. (2006). Atmospheric sounding over the winter Kuroshio Extension: Effect of surface stability on atmospheric boundary layer structure. *Geophysical Research Letters*, 33, L04703. https://doi.org/10.1029/2005GL025102

Trenberth, K. E. (1986). An assessment of the impact of transient eddies on the zonal flow during a blocking episode using localized Eliassen-Palm flux diagnostics. *Journal of the Atmospheric Sciences*, 43(19), 2070–2087. https://doi.org/10.1175/1520-0469(1986)043<2070:AAITOE>2.0.CO;2

Trenberth, K. E. (1984). Interannual variability of the Southern Hemisphere circulation: Representativeness of the year of the global weather experiment. *Monthly Weather Review*, 112(1), 108–123. https://doi.org/10.1175/1520-0493(1984)112<0108:IVSTSH>2.0.CO;2

Vecchi, G. A., Xie, S. P., & Fischer, A. S. (2004). Ocean–atmosphere covariability in the Western Arabian Sea. *Journal of Climate*, 17(6), 1213–1224. https://doi.org/10.1175/1520-0442(2004)017<1213:OACITW>2.0.CO;2

Wallace, J. M., Smith, C. A., & Jiang, Q. (1990). Spatial patterns of atmosphere-ocean interaction in the northern winter. *Journal of Climate*, 3(9), 990–998. https://doi.org/10.1175/1520-0442(1990)003<0990:SOAOIT>2.0.CO;2

Wang, L. Y., Hu, H. B., & Yang, X. Q. (2019). The atmospheric responses to the intensity variability of subtropical front in the wintertime North Pacific. *Climate Dynamics*, 52(9–10), 5623–5639. https://doi.org/10.1007/s00382-018-4468-9

Wang, L. Y., Hu, H. B., Yang, X. Q., & Ren, X. J. (2016). Atmospheric eddy anomalies associated with the wintertime North Pacific subtropical front and their influences on the seasonal-mean atmosphere. *Science China: Earth Sciences*, 59(10), 2022–2036. https://doi.org/10.1007/s11430-016-5331-7

Wang, Z. Y., Song, Z. Y., & Liu, G. L. (2020). The characteristics of near-equatorial North Pacific low PV water and its possible influences on the equatorial subsurface ocean. *Journal of Geophysical Research: Oceans*. 125, e2020JC016282. https://doi.org/10.1029/2020JC016282

Wang, L. Y., Yang, X. Q., Yang, D. J., Xie, Q., Fang, J. B., & Sun, X. G. (2017). Two typical modes in the variabilities of wintertime North Pacific basin-scale oceanic fronts and associated atmospheric eddy-driven jet. *Atmospheric Science Letters*, 18(9), 373–380. https://doi.org/10.1002/asl.766

Wen, Z. B., Hu, H. B., Song, Z., Bai, H. K., & Wang, Z. (2020). Different influences of mesoscale oceanic eddies on the north pacific subsurface low potential vorticity water mass between winter and summer. *Journal of Geophysical Research: Oceans*, 125, https://doi.org/10.1029/2019JC015333

Xiao, C. L., & Zhang, Y. C. (2012). The East Asian upper-tropospheric jet streams and associated transient eddy activities simulated by a climate system model BCC-CSM1.1. *Journal of Meteorological Research*, 26(6), 700–716. https://doi.org/10.1007/s13351-012-0603-4

Xue, Y., Juang, H. M., Li, W., Prince, S., DeFries, R., Jiao, Y., & Vasic, R. (2004). Role of land surface processes in monsoon development: East Asia and West Africa. *Journal of Geophysical Research*, 109, D03105. https://doi.org/10.1029/2003JD003556

Xu, H. M., Xu, M. M., Xie, S. P., & Wang, Y. Q. (2011). Deep atmospheric response to the spring Kuroshio over the East China Sea. *Journal of Climate*, 24(18), 4959–4972. https://doi.org/10.1175/JCLI-D-10-05034.1

Xu, M. M., & Xu, H. M. (2015). Atmospheric responses to Kuroshio SST front in the East China Sea under different prevailing winds in winter and spring. *Journal of Climate*, 28(13), 3191–3211. https://doi.org/10.1175/JCLI-D-13-06457.1

Xue, Y., Jiang, H. M., Li, W., Prince, S., DeFries, R., Jiao, Y., & Vasic, R. (2004). Role of land surface processes in monsoon development: East Asia and West Africa. *Journal of Geophysical Research*, 109, D03105. https://doi.org/10.1029/2003JD003556

Yao, Y., Zhong, Z., & Yang, X. Q. (2016). Numerical experiments of the storm track sensitivity to oceanic frontal strength within the Kuroshio/Oyashio extensions. *Journal of Geophysical Research, 121*, 2888–2900. https://doi.org/10.1002/2015JD024381

Zhang, C., Liu, H. L., Liu, C. Y., & Lin, P. F. (2019). Impacts of mesoscale sea surface temperature anomalies on the meridional shift of North Pacific storm track. *International Journal of Climatology*, 39, 5124–5139. https://doi.org/10.1002/joc.6130

Zhang, L. Y., Xu, H. M., Shi, N., & Deng, J. C. (2017). Responses of the East Asian jet stream to the North Pacific subtropical front in spring. *Advances in Atmospheric Sciences*, 34(2), 144–156. https://doi.org/10.1007/s00376-016-6026-x

Zhang, Y. C., & Huang, D. Q. (2011). Has the East Asian westerly jet experienced a poleward displacement in recent decades. *Advances in Atmospheric Sciences*, 28(6), 1259–1265. https://doi.org/10.1007/s00376-011-9183-9

Zhang, Y. Z., Yan, P. W., Liao, Z. J., Huang, D. Q., & Zhang, Y. C. (2019). The winter concurrent meridional shift of the East Asian jet streams and the associated thermal conditions. *Journal of Climate*, 32(7), 2075–2088. https://doi.org/10.1175/JCLI-D-18-0085.1

Zhang, Y., Yang, X.-Q., Nie, Y., & Chen, G. (2012). Annual mode-like variation in a multilayer quasigeostrophic model. *Journal of the Atmospheric Sciences*, 69(10), 2940–2958. https://doi.org/10.1175/JAS-D-11-0214.1

Zhou, T. J., & Yu, R. C. (2005). Atmospheric water vapor transport associated with typical anomalous summer rainfall patterns in China. *Journal of Geophysical Research*, 110, D08104. https://doi.org/10.1029/2004JD005413

Zhou, T. J., & Zou, L. (2010). Understanding the predictability of East Asian summer monsoon from the reproduction of land-sea thermal contrast change in AMIP-type simulation. *Journal of Climate*, 23(22), 6009–6026. https://doi.org/10.1175/2010JCLI3546.1

Zhu, Z. W., & Li, T. (2016). A new paradigm for continental U.S. summer rainfall variability: Asia-North America teleconnection. *Journal of Climate*, 29(20), 7313–7327. https://doi.org/10.1175/JCLI-D-16-0137.1