Divergent Responses of Extratropical Atmospheric Circulation to Interhemispheric Dipolar SST Forcing over the Two Hemispheres in Boreal Winter

JIAQING XUE, a,b CHENG SUN, c JIANPING LI, c,d JIANGYU MAO, a HISASHI NAKAMURA, e TAKAFUMI MIYASAKA, e,f AND YIDAN XU c

a State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China
b College of Earth Science, University of Chinese Academy of Sciences, Beijing, China
c College of Global Change and Earth System Science, Beijing Normal University, Beijing, China
d Laboratory for Regional Oceanography and Numerical Modeling, Qingdao National Laboratory for Marine Science and Technology, Qingdao, China
e Climate Science Research Laboratory, Research Center for Advanced Science and Technology, The University of Tokyo, Tokyo, Japan
f Meteorological Research Institute, Japan Meteorological Agency, Tsukuba, Japan

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ABSTRACT

Global sea surface temperature (SST) evolution exhibits an antiphase variation between the two hemispheres that is referred to as the SST interhemispheric dipole (SSTID) mode. The impacts of the SSTID on extratropical atmospheric circulation in boreal winter are explored by both regression analysis and SST-forced numerical simulations. The responses of extratropical circulation to SSTID thermal forcing bear an equivalent barotropic structure. For the Southern Hemisphere (SH), positive SSTID events lead to a meridional dipolar perturbation in sea level pressure (SLP), similar in pattern to the positive southern annular mode (SAM). Although SSTID-forced SLP anomalies over the Northern Hemisphere (NH) do not exhibit a zonally symmetric pattern as is the case over the SH, they still show signs of a meridional dipole opposite to the SH over the oceans. Divergent circulation responses to SSTID forcing between the two hemispheres are suggested to be associated with contrasting storm-track variations. Positive SSTID events weaken oceanic fronts in both the North Atlantic and North Pacific, and thus lead to the decline of NH storm-track activity by decreasing atmospheric baroclinicity. In the SH, positive SSTID events correspond to the enhancement of SH transients by intensifying the Antarctic polar-frontal zone. Additionally, local baroclinic energy conversions are diagnosed to explain the SSTID-related storm-track variations over both hemispheres. Finally, an investigation of transient eddy feedback indicates that the SSTID mode modulates extratropical atmospheric circulation, primarily by regulating storm tracks and changing the corresponding eddy feedback.

1. Introduction

Aside from first-order global anthropogenic warming, surface temperature exhibits prominent nonuniform variations across different latitudes (Braganza et al. 2003; Sutton et al. 2007; Drost et al. 2012). A noticeable manifestation of these nonuniform variations is that the warming rates of surface temperature between the two hemispheres are unsynchronized, with the warming of Northern Hemisphere (NH) being faster than that of Southern Hemisphere (SH) for one subperiod, and the opposite being true for another (Drost et al. 2012; Xu and Ramanathan 2012). An emerging indicator, defined as the difference between hemispheric-mean surface temperatures, has been devised to measure the temperature asymmetry between the two hemispheres (Drost et al. 2012; Friedman et al. 2013).

Additionally, the interhemispheric asymmetric evolution has been revealed to exist in the sea surface temperature (SST) field. Sun et al. (2013) investigated global SST variability by applying empirical orthogonal function (EOF) analysis to the zonally averaged SST anomalies. They found, aside from the global warming signal, that zonal-mean SST variability is dominated by an equatorial symmetric mode and an equatorial antisymmetric mode. The symmetric mode is shown to be associated with the interdecadal Pacific oscillation (IPO), whereas the antisymmetric mode is characterized...
by opposite-signed SST anomalies between the two hemispheres, and is referred to as the SST interhemispheric dipole (SSTID) mode. NH SST warming with concurrent SH SST cooling is defined as positive phases of the SSTID. A similar SST variability of such interhemispheric antisymmetric pattern has also been documented in other previous studies (Folland et al. 1999; Cai and Whetton 2001; Dima and Lohmann 2010; Thompson et al. 2010; Wang et al. 2015). In one such study, by directly applying EOF analysis to global SST anomalies, Dima and Lohmann (2010) identified an interhemispheric seesaw pattern in SST variability, which bears a strong resemblance to the SSTID. Although the most prominent feature of the SSTID is the interhemispheric seesaw of SSTs, SSTID-related SST anomalies are also nonuniform in each hemisphere with the largest amplitude appearing at the middle to high latitudes (Dima and Lohmann 2010; Thompson et al. 2010; Chiang and Friedman 2012).

Storm tracks are indicative of preferred occurrence regions for synoptic-scale transient eddies in the mid-latitudes of both hemispheres (Blackmon et al. 1977; Chang et al. 2002). High-frequency baroclinic transients migrating along the storm tracks not only impact day-to-day weather, but can also modulate planetary-scale mean flows through the eddy feedback (Lau and Holopainen 1984; Cai and Mak 1990; Jin et al. 2006). Climatologically, transient eddies participate in global energy and hydrological cycles by transporting heat and water vapor poleward. Meanwhile, the meridional eddy flux of zonal momentum also plays a critical role in maintaining the midlatitude jets and keeping Earth’s angular momentum budget in balance (Hartmann 1994). Because storm-track regions are also areas of intense eddy–mean-flow interactions, any systematic changes in the strength or location of storm tracks can have remarkable impacts on large-scale atmospheric circulation and climate variability (Rogers 1997; Lau and Nath 1991; Salathé 2006).

Vigorous transient eddy activity in storm-track regions is due primarily to the baroclinic instability of mean flows over these areas (Charney 1947; Eady 1949). Because meridional heat transport by transient eddies, to some degree, tends to relax the baroclinicity of mean flows that sets up storm tracks in the first place, the maintenance of storm tracks needs a coupling with underlying oceans to keep the baroclinicity of mean flows. Therefore, the climatological distribution of storm tracks is anchored by midlatitude oceanic fronts in the North Pacific, North Atlantic, and Southern Oceans through setting baroclinic zones in the atmosphere (Nakamura et al. 2004; Nakamura and Shimpo 2004; Shaw et al. 2016). The intimate linkages between storm tracks and underlying SST thermal forcing indicate that extratropical SST anomalies can systematically modulate high-frequency transients and thus change large-scale atmospheric circulation through the eddy feedback (Peng and Whitaker 1999; Kushnir et al. 2002; Brayshaw et al. 2008; Nie et al. 2016).

As introduced above, SST anomalies associated with the SSTID are nonuniform in each hemisphere, so SSTID variability may be accompanied by changes in midlatitude SST gradients. Meanwhile, storm-track activity is closely linked to underlying oceanic frontal zones. Therefore, the SSTID mode has the potential to modulate extratropical large-scale circulation by systematically changing transient eddies and the corresponding eddy feedback.

Nevertheless, on the climate impacts of interhemispheric dipolar SST forcing, previous studies concentrated primarily on the tropics from both observational and theoretical perspectives (Folland et al. 1986; Kang et al. 2008; Chiang and Friedman 2012; Sun et al. 2013; Schneider et al. 2014; Lopez et al. 2016). In contrast to detailed understandings of tropical responses to the SSTID, possible influences of interhemispheric dipolar SST forcing on extratropical atmospheric variability remain unclear, and further investigations into this topic are needed.

The main concern of present study is to investigate the SSTID-related changes in extratropical circulation, storm-track activity as well as the feedback between them. The remainder of this paper is organized as follows. Datasets and diagnostic methods are introduced in section 2. Section 3 describes basic spatiotemporal features of the SSTID mode. The influences of the SSTID on extratropical atmospheric circulation and corresponding physical explanations in terms of the transient eddy feedback are investigated in section 4. Finally, section 5 contains a summary and discussion.

2. Datasets and methodology

a. Data and index definitions

Gridded SST data covering the period 1900–2010 are primarily based on the Extended Reconstructed Sea Surface Temperature version 4 (ERSSTV4), which adopts new bias adjustments and quality control procedures compared with version 3b (Huang et al. 2015). In addition, SST data taken from the Hadley Centre Sea Ice and Sea Surface Temperature dataset version 2 (HadISST2) are employed to test reliability of the results (Kennedy et al. 2013; Poli et al. 2016). Sea ice concentration is also obtained from the HadISST2 dataset (Titchner and Rayner 2014). Monthly-mean and daily atmospheric reanalysis data, including air
temperature, sea level pressure (SLP), geopotential height, and 3D wind fields, are derived primarily from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) dataset (Kalnay et al. 1996), covering the period 1948–2010. Moreover, monthly-mean circulation fields from the European Centre for Medium-Range Weather Forecast twentieth-century reanalysis (ERA-20C) (Poli et al. 2016) for a longer period 1900–2010 are also employed for parallel analysis.

The SST-forced ensemble simulations based on the NASA GISS Model E2-R are obtained from Atmospheric Model Intercomparison Project (AMIP) experiment in phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012). This model is selected for its availability of long-term simulations, which make it possible to obtain reliable atmospheric responses to boundary SST forcing. These AMIP simulations are run with prescribed historical SST for the period 1880–2010 and include six ensemble members starting from different initial conditions (r1i1p1–r6i1p1). Unlike an individual model simulation that contains atmospheric internal variability, the ensemble mean of a set of atmospheric general circulation model (AGCM) runs can be utilized to limit internal variability and highlight SST-forced atmospheric responses. Therefore, the analysis in the present study is based on the six-member ensemble mean of these AMIP simulations. Moreover, the first 20 years of all simulations are regarded as spinup and the analysis is performed on the remaining period 1900–2010.

The SSTID index used in this study is directly constructed as the difference of hemispheric-mean SST anomalies between the two hemispheres (i.e., NH minus SH) for more explicit physical meaning, which is highly consistent with the definition based on the EOF analysis (Sun et al. 2013). The IPO index is calculated as the difference between SST anomalies averaged over the central equatorial Pacific (10°S–10°N, 170°E–90°W) and the average of SST anomalies in the northwest Pacific (25°–45°N, 140°E–145°W) and southwest Pacific (50°–15°S, 150°E–160°W) following Henley et al. (2015). The Atlantic multidecadal oscillation (AMO) index is defined as the area-weighted average of SST anomalies over the North Atlantic region (0°–65°N) (Enfield et al. 2001). The southern annular mode (SAM) index here is defined as the difference in the normalized zonal-mean SLP between 40° and 70°S (Nan and Li 2003), which is a modification of the Antarctic Oscillation (AAO) index defined by Gong and Wang (1999). All anomalous variables used in this study are calculated with respect to the 1960–90 climatology. Winter in this study refers to the boreal season and is defined as December, January, and February (DJF).

b. Analysis tools

Storm-track characteristics are depicted with the bandpass-filtered eddy statistics (variance or covariance), which isolate transient disturbances by applying a bandpass filter (2–7 days) to the daily time series at each grid point (Blackmon et al. 1977). Therefore, storm tracks in this study are characterized primarily by the root-mean-square (RMS) of the 2–7-day bandpass-filtered geopotential height \( \bar{Z}^{2/2} \), where the prime indicates high-frequency transient quantity and the overbar represents time averaging over an individual month. Moreover, meridional eddy heat transport \( \overline{v' T} \), eddy flux of westerly momentum \( \overline{u' v} \), and wave energy dispersion by storm tracks are also examined. The direction of wave energy propagation relative to the time-mean flow is pointed by the extended Eliassen–Palm vectors (hereafter \( \mathbf{E} \) vectors), which can also be used to diagnose the dynamical forcing of transient eddies on the mean flow (Hoskins et al. 1983). In this study, horizontal components of \( \mathbf{E} \) vectors are defined by Trenberth (1986)

\[
\frac{1}{2} (u'^2 - u^2) i - u' v' j,
\]  

(1)

where \( u \) and \( v \) represent zonal and meridional winds, respectively. The divergence (convergence) of \( \mathbf{E} \) vectors relates to the eddy-induced accelerations (decelerations) of mean zonal winds.

The feedback forcing of transient eddies on time-mean flows is evaluated by the eddy-induced geopotential height tendency. Following Lau and Holopainen (1984), the height tendency forced by transient eddy vorticity and heat fluxes can be represented by a 3D quasigustostrophic potential vorticity equation:

\[
\begin{align*}
\nabla^2 \psi + f^2 \frac{\partial}{\partial p} \left( \frac{1}{S} \frac{\partial Z}{\partial t} \right) & = -\frac{g}{\mathcal{F}} \nabla \cdot \left( \nabla \zeta \right) \\
& + \frac{f^2}{g} \frac{\partial}{\partial p} \left[ \nabla \cdot \left( \frac{V' \theta'}{\mathcal{F}} \right) \right].
\end{align*}
\]  

(2)

Here the prime denotes high-frequency transient quantities and the overbar indicates time averaging. Also, \( \partial Z / \partial t \) represents the low-frequency geopotential height tendency, \( S = -\partial \Theta / \partial p \) is the static stability parameter, \( f \) denotes the Coriolis parameter, \( \Theta \) is the potential temperature of the background state, \( V \) is the horizontal wind vector, \( \zeta \) represents the relative vorticity, and \( \theta \) is the potential temperature (Nishii et al. 2009; Tanaka et al. 2016). The first term on right-hand side of Eq. (2) is
connected to eddy vorticity flux (i.e., the eddy vorticity forcing) and the second term is associated with eddy heat flux (i.e., the eddy thermal forcing). As described in Lau and Holopainen (1984), the geopotential height tendency $\partial Z/\partial t$ can be obtained by solving the 3D potential vorticity equation with relaxation method under appropriate boundary conditions.

Baroclinic instability of mean flows is the key dynamic mechanism responsible for active transient eddy activity in storm-track regions (Charney 1947; Eady 1949). The maximum Eady growth rate (defined as $\sigma = 0.31gN^{-1}T^{-1}|\partial T/\partial y|$, where $T$ is temperature and $N$ denotes the Brunt–Väisälä frequency) can be diagnosed to measure atmospheric baroclinicity and thus indicate storm-track activity (Lindzen and Farrell 1980). Nevertheless, various factors can modulate the efficiency of eddies’ ability to tap into the baroclinicity of mean flows, given that energy conversions from available potential energy to eddy kinetic energy play a dominant role in maintaining the transients. Therefore, an investigation of storm-track variability from the perspective of local baroclinic energy conversions is warranted and has been widely adopted in a range of previous studies (Lee et al. 2011, 2012; Gan and Wu 2015). The baroclinic energy conversion (BCEC) processes include conversions from mean available potential energy (MAPE) to eddy available potential (EAPE) and then from EAPE to eddy kinetic energy (EKE), which can be expressed as follows:

$$\text{BCEC(MAPE} \rightarrow \text{EAPE)} = -C_1 \frac{p_0}{p} \frac{R_C}{C_p} \left( -\frac{\partial \theta}{\partial p} \right)^{-1} \left( \frac{\vec{u} \cdot \vec{T}}{\partial x} + \frac{\vec{v} \cdot \vec{T}}{\partial y} \right),$$

(3)

$$\text{BCEC(EAPE} \rightarrow \text{EKE)} = -C_1 \left( \vec{\omega} \vec{T} \right).$$

(4)

The prime here indicates synoptic-scale transients, while the overbar signifies averaging over an individual month, and $C_1 = (p_0/p)C_0/G^2 R/g$. Moreover, $C_0$ ($C_p$) is the specific heat of dry air at the constant pressure (volume); $R$, $g$, and $\theta$ represent the gas constant for dry air, acceleration of gravity, and potential temperature, respectively. The mean sea level pressure $p_0$ is taken as 1000hPa. More detailed derivations can be found in Cai et al. (2007).

3. Temporal and spatial features of the SSTID mode

The normalized annual SSTID indices during the period 1900–2010 are presented in Fig. 1a, and the indices computed from two sets of SST data are well correlated ($r = 0.88$, significant at the 95% confidence level). The SSTID index varies on both interannual and interdecadal time scales and shows an abrupt change from a positive phase to a negative phase around 1970. The spatial pattern of the SSTID then can be obtained by regressing global SST onto the normalized SSTID index. As shown in Fig. 1b, the SSTID mode is characterized by prominent interhemispheric dipolar pattern with opposite-signed SST anomalies between the two hemispheres. Moreover, it is noted that the SSTID signature is nonuniform in each hemisphere with the largest SST anomalies appearing at the middle to high latitudes. Although the exact formation mechanisms for the SSTID are still not well understood, both oceanic meridional overturning circulation (MOC) (Latif et al. 2006; Lopez et al. 2016) and hemispheric asymmetries in aerosol emissions (Tett et al. 2002; Chung and Soden 2017) may contribute to the asymmetric evolution of SST between the two hemispheres. Modeling results demonstrate that SST anomalies related to MOC variations and anthropogenic aerosol forcing show the largest loadings in subpolar regions (Zhang et al. 2013; Lopez et al. 2016; Chung and Soden 2017), which may explain why SST anomalies related to the SSTID are nonuniform in each hemisphere and peak at the mid- to high-latitude regions. This salient feature of the SSTID mode can be more clearly seen from the regressed pattern of zonal-mean SST anomalies shown on the right panel of Fig. 1b. In the NH, positive SST anomalies are observed to peak at the latitudinal band of 50°–60°N, while the strongest negative SST anomalies are found at 60°S in the SH. The SSTID pattern computed from the HadISST2 dataset is further presented in Fig. 1c, which bears a considerable resemblance to that based on the ERSST data, indicating that the SSTID pattern is independent of the choice of SST datasets.

Explicit trend analysis is further employed to demonstrate the antiphase evolution of SST between the two hemispheres. Figure 2 presents linear trends of residual SST field, from which both global warming and IPO signals have been linearly removed. The global warming signal here is represented by the globally averaged SST anomalies, as suggested by Trenberth and Shea (2006). The periods 1910–35 and 1960–80, both corresponding to major phase transitions of the SSTID index, are selected for analysis. During the period 1910–35, NH SST was dominated by notable warming over both the Pacific and Atlantic basins, whereas in the entire SH, especially the Southern Ocean region, significant cooling occurred in sharp contrast to the NH (Figs. 2a,c). During the period 1960–80, the entire NH experienced significant cooling in comparison with hemispheric-wide SST warming in the SH (Figs. 2b,d). Moreover, an inspection of zonal-mean SST trend patterns also reveals that SSTID-related
SST variations are most prominent over the middle to high latitudes.

Although the above spatiotemporal features of the SSTID are revealed in an annual-mean sense, the SSTID mode is found to be prominent in all four seasons (not shown). Because storm-track activity and the corresponding transient eddy feedback are active over both hemispheres in boreal winter (Trenberth 1991; Chang et al. 2002; Nakamura and Shimpo 2004), the DJF is thus selected to be the research period for this study. Figure 1d further presents the SSTID pattern in DJF, and it is found that the SSTID pattern in boreal winter shows similar spatial distributions to the annual-mean situation.

4. Signatures of the SSTID on wintertime atmospheric variability over the extratropics

a. Extratropical circulation responses to SSTID thermal forcing

The association of wintertime atmospheric circulation with SSTID variability can be determined by regressing anomalous circulation variables onto the normalized SSTID index. Because the SSTID pattern demonstrates a higher degree of zonal symmetry in the SH, we first examine the SSTID-related atmospheric pattern over the SH. Figure 3a shows the regression map of DJF-mean SLP anomalies over the SH based on the NCEP reanalysis data. The anomalous SLP features a significant meridional dipolar pattern with zonally symmetric structure. This annular pattern corresponds to substantially increased SLP to the north of 60°S with concurrent decreases in SLP to the south, which is similar to the positive SAM pattern (Li et al. 2016; Liu et al. 2017; Zheng et al. 2018). So, the relationship between the SSTID and SAM variability in boreal winter is further examined. The variation of the SAM index is found to be basically in phase with the SSTID index in DJF (not shown). The correlation coefficient between them is 0.3, which is significant at the 95% confidence level.

Because of different land–sea distribution, the SSTID pattern in the NH is zonally asymmetric. Therefore, in the NH, we primarily focus on anomalous atmospheric variability over the oceans (i.e., the North Pacific and North Atlantic regions). As shown in Fig. 3b, over the North Pacific, the anomalous SLP distribution is characterized by positive values near the Aleutian Islands and weak negative anomalies to the south. For the North Atlantic basin, SLP anomalies are dominated by a significant north–south-oriented dipolar pattern with
increased SLP over southern Greenland and decreased SLP to the south spanning from the Caribbean Sea to northern Europe, which is reminiscent of the negative North Atlantic Oscillation (NAO) (Li and Wang 2003). Overall, the anomalous NH SLP does not exhibit an annular meridional dipole as is the case over the SH, due to the presence of weak low pressure anomalies over the North American continent. However, when examining the patterns over the two ocean basins separately, the anomalous SLP patterns retain a generally north–south dipolar structure, particularly over the North Atlantic region. Parallel analyses employing the ERA-20C dataset are further presented for comparison in the right panel of Fig. 3, which bears considerable resemblance to the patterns derived from the NCEP dataset, indicating the robustness of the SSTID-related circulation patterns.

The anomalous 300-hPa geopotential height linked to the SSTID is also examined to reveal the vertical structure of SSTID-related changes in extratropical circulation. As demonstrated in Fig. 4, the 300-hPa regression pattern in the SH also exhibits an annular pattern with increased geopotential height to the north of 60°S and decreased geopotential height to the south (Figs. 4a,c). Moreover, over extratropical regions in the NH, the regressed 300-hPa geopotential height features positive anomalies to the south of both the Aleutian Islands and Greenland, while negative anomalies are found west of Hawaii and in a striped domain extending from Florida across the North Atlantic into northern Europe (Figs. 4b,d). A comparison between Figs. 3 and 4 reveals that anomalous extratropical circulations related to the SSTID exhibit an equivalent barotropic structure.

The SSTID-related anomalous extratropical circulations are dominated by meridional dipolar structures over both hemispheres, so SSTID variability should be accompanied by significant changes in midlatitude zonal winds. Because of the impacts of land–sea distribution, anomalous circulation in the NH shows strong zonal asymmetry, so Figs. 5a–c investigate SSTID-related changes in westerly jets over the SH, North Atlantic, and the North Pacific, respectively. In association with the SSTID, strengthened westerly winds in the SH are located to the south of climatological midlatitude jet center, indicating a poleward shift of the westerly jet (Fig. 5a). Therefore, the SSTID-related positive SAM-like pattern leads to a latitudinal shift of westerly jet in the SH, consistent with the results in previous studies (Lorenz and Hartmann 2001; Nie et al. 2013). The regressed zonal winds over the North Atlantic sector feature a meridional dipole straddling the climatological jet center, which indicates the equatorward shift of Atlantic midlatitude jet (Fig. 5b). For the North Pacific
region, the climatological midlatitude jet center denoted by near-surface wind maximum is located near 40°N, while the SSTID-related weakening of zonal winds is located over the latitudinal band of 30°–40°N, which also signifies a latitudinal shift of the Pacific jet center (Fig. 5c).

By regression analysis above, we find that SSTID variability is accompanied by distinct changes in extratropical circulation over the two hemispheres. Whether such anomalous circulation is forced by SSTID-related dipolar thermal forcing is further examined by SST-forced ensemble simulations in the AMIP experiment. Figure 6 presents large-scale circulation responses to SSTID forcing in AMIP simulations by regressing both SLP and 300-hPa geopotential height onto the normalized SSTID index. In the SH, the responses of SLP display an annular pattern with increased SLP to the north of 60°S and decreased SLP over the polar region, which is analogous to the positive SAM pattern (Fig. 6a). For the NH, SLP responses are characterized by positive anomalies near the Aleutian Islands, and a significant north–south-oriented dipole over the North Atlantic with increased SLP over the southern Greenland and decreased SLP to the south (Fig. 6b). Figures 6c and 6d further display 300-hPa geopotential height responses to the SSTID in AGCM simulations, which are similar in pattern to that of SLP over the extratropical regions, indicating equivalent barotropic responses. Comparing Fig. 6 with Figs. 3 and 4, it is found that the SSTID-related circulation patterns identified by regression analysis can be well reproduced by AGCM simulations forced by SSTID forcing, which demonstrates that the SSTID can indeed have divergent modulating impacts on extratropical circulation over the two hemispheres in boreal winter.

The transient eddy feedback plays a fundamental role in determining atmospheric responses to underlying SST forcing over the extratropics. To explore underlying dynamical mechanisms responsible for the observed responses of extratropical circulation to the SSTID, changes in oceanic fronts, storm-track activity, and the corresponding eddy feedback accompanying the SSTID are further examined.

b. The relationship between the SSTID and oceanic frontal zones

Figure 7a depicts the climatological collocation of 300-hPa storm tracks (shading) and midlatitude oceanic fronts (contours) in boreal winter. Oceanic fronts
characterized by intense meridional SST gradients include the North Pacific subarctic frontal zone in the Kuroshio–Oyashio Extension, the North Atlantic subarctic front in the Gulf Stream Extension, and the Antarctic polar-frontal zone in the Southern Ocean, consistent with oceanic frontal zones identified in previous studies (Orsi et al. 1995; Nakamura et al. 2004; Marshall and Speer 2012). Moreover, active storm tracks are observed to be over the extratropical North Pacific and North Atlantic in the NH, and over the Southern Ocean basin in the SH. Further inspection of Fig. 7a reveals that storm-track distributions over the two hemispheres coincide well with oceanic fronts except for slightly poleward shift, suggesting an intimate linkage between storm-track activity and underlying SST thermal forcing.
During positive phases of the SSTID, the largest SST warming is located over the midlatitude North Pacific and the subpolar North Atlantic in the NH, while the strongest SST cooling appears over the Southern Ocean in the SH (Fig. 1d), so it can be inferred that positive SSTID events tend to decrease (increase) the intensity of meridional SST gradients in NH (SH) oceanic frontal zones. To confirm this, Fig. 7b presents the correlation map between the SSTID index and the intensity of meridional SST gradients in boreal winter, with climatological ocean fronts shown as contours. In association with positive SSTID events, the intensity of subtropical oceanic front is decreased over the North Pacific; meanwhile, the decreasing of SST front in the North Atlantic is located on the south side of the climatological oceanic front. For the SH, the Antarctic polar-frontal zone in the Southern Ocean is strengthened in response to negative subpolar SST anomalies.

Except for the SSTID, dominant modes of SST variability in extratropical oceans include the IPO and AMO. Whether contributions of the SSTID to oceanic fronts are affected by these two SST modes is further examined by partial correlation analysis. Figure 7c shows the partial correlation between SST gradient intensity and the SSTID index after removing the IPO signal. Comparing Fig. 7c with Fig. 7b, it is found that the relationship between the SSTID and oceanic fronts remains nearly unchanged with the IPO signal removed. This is because the IPO as an equatorial symmetric mode is independent of the SSTID, and the SSTID and IPO indices are nearly uncorrelated. Similarly, Fig. 7d further presents the partial correlation between the SSTID index and the intensity of SST gradients excluding the AMO signal. For the North Atlantic, the relation between the SSTID and meridional SST gradients is slightly weakened after removing the AMO signal; for other basins outside the North Atlantic, the contributions of the SSTID to oceanic fronts are unaffected by the AMO. The AMO as a regional SST mode over the North Atlantic is asymmetric about the equator, so the AMO signal will be included when defining the SSTID index. Therefore, the SSTID signal in North Atlantic SST is partly contributed by the AMO, which may explain why the relation between the SSTID and meridional SST gradients over the North Atlantic is weakened after removing the AMO. However, even with the AMO signal removed, the influences of the SSTID on oceanic fronts over the North Atlantic is still significant. Based on the above analysis, it is demonstrated that the contributions of SSTID variability to

**Fig. 6.** (a) Regression of DJF-mean SH SLP anomalies (units: hPa) against the normalized SSTID index during the period 1900–2010 in SST-forced (AMIP) ensemble simulations. (b) As in (a), but for the North Pacific and North Atlantic regions. (c),(d) As in (a),(b), but for 300-hPa geopotential height anomalies (units: gpm). Dotted shading denotes regressions significant at the 90% confidence level (Student’s $t$ test).
oceanic fronts are largely independent of other modes of SST variability.

c. Divergent storm-track variations related to SSTID over the two hemispheres

The above results indicate that the SSTID mode can exert opposite impacts on oceanic fronts over the two hemispheres. Meanwhile, based on numerical experiments, many previous studies have elucidated the important roles of oceanic fronts in anchoring atmospheric baroclinic zones and thus maintaining storm tracks via differential heat supply from the oceans (Nakamura et al. 2008; Nonaka et al. 2009; Taguchi et al. 2009; Sampe et al. 2010; Hotta and Nakamura 2011). So, the SSTID-related fluctuations in oceanic fronts should be accompanied by changes in atmospheric baroclinicity and storm tracks. The atmospheric baroclinicity is measured by the maximum Eady growth rate, which is closely linked to meridional air temperature gradients. Because anomalous air temperature acts as a bridge to link SSTID-related SST anomalies and changes in atmospheric baroclinicity, air temperature variations related to the SSTID are first examined. Figure 8a depicts the regressed pattern of air temperature averaged over the North Atlantic sector with corresponding climatological air temperature shown as contours. Corresponding to strongest positive SST anomalies in the subpolar North Atlantic, SSTID-related maximum tropospheric warming is found to be over the latitudinal band of 50°–60°N. In response to this anomalous warming, atmospheric baroclinicity over the North Atlantic weakens over the region of 40°–50°N (Fig. 8b). The collocation of regressed Eady growth rate and its climatological profile indicates the weakening of atmospheric baroclinicity over the North Atlantic. For the North Pacific region, maximum tropospheric warming appears over the strongest positive SST anomalies near 45°N, thus reducing atmospheric baroclinicity over the region of 30°–45°N (Figs. 8c,d) and also indicating a weakening of atmospheric baroclinicity over the North Pacific. Moreover, SSTID-related changes in air temperature and the maximum Eady growth rate over the SH are examined in Figs. 8e and 8f. In contrast to the circumstance in the NH, negative SST anomalies in the Southern Ocean lead to pronounced tropospheric cooling south of 60°S. In conjunction with this cooling pattern,
atmospheric baroclinicity strengthens over the region of 50°–60°S, which is located to the south flank of climatological peak. Therefore, positive SSTID events correspond to a poleward shift of atmospheric baroclinic zones in the SH (Fig. 8f).

Following the above analysis on vertical structures of anomalous Eady growth rate, Fig. 9a further explores the horizontal distribution of SSTID-related changes in the NH Eady growth rate. Over the North Pacific, the regressed 700-hPa Eady growth rate is characterized by negative anomalies located to the south of the climatological peak, whereas for the North Atlantic region decreased Eady growth rate is observed on the poleward flank of the climatological maxima. Prominent decreases in NH Eady growth rate indicate that the storm-track system may undergo systematic changes.
accompanied by the SSTI. As shown in Fig. 9b, the climatological 300-hPa $Z^{21/2}$ in the NH is dominated by two maxima centered over the North Pacific and North Atlantic sectors, corresponding to NH storm-track regions. Consistent with the regressed pattern of the maximum Eady growth rate, significant negative anomalies of $Z^{21/2}$ are located to the south flank of the climatological peak over the North Pacific, signifying a weakening and slightly poleward shift of the Pacific storm track. For the North Atlantic sector, the regressed $Z^{21/2}$ exhibits prominent negative anomalies on the northern part of climatological storm track, indicating an equatorward shift of the North Atlantic storm track (Fig. 9b).

Moreover, signatures of storm-track variations on the meridional eddy heat flux ($\nabla T\overline{v}$), eddy momentum flux ($\overline{u\nabla v}$), and wave energy dispersion (E vectors) are further examined. The regression field of 700-hPa meridional eddy heat transport due to weakened storm-track activity. Transient eddies also play a fundamental role in maintaining the midlatitude westerly jet against dissipation at the surface through the convergence of westerly momentum. The regressed pattern of eddy momentum flux in the NH is shown as shading in Fig. 9d, with positive (negative) anomalies located to the north (south) of storm-track regions, in contrast to the climatological pattern of eddy westerly momentum flux, signifying decreased convergence of eddy westerly momentum due to weakened storm tracks in the NH. Meanwhile, accompanying the weakened transient eddy activity, anomalous wave energy is observed to converge into NH storm-track regions, and the convergence of E vectors indicates the deceleration of westerly mean flows.

Analogous to analyses for the NH, SSTI-related changes in storm-track characteristics over the SH are presented in Fig. 10. A significantly increased low-level Eady growth rate appears over the latitudinal band of 50°–60°S, indicating possible strengthened SH storm tracks (Fig. 10a). To confirm this, Fig. 10b gives the regressed pattern of 300-hPa $Z^{21/2}$, which exhibits significant positive values centered to the south side of climatological $Z^{21/2}$, indicating poleward enhancement of SH storm tracks (Fig. 10b). The manifestations of intensified storm tracks in $\nabla T\overline{v}$, $\overline{u\nabla v}$, and the E vectors are examined in Figs. 10c and 10d. The regression pattern of $\nabla T\overline{v}$ in the lower troposphere is characterized by
anomalous southward heat transport superimposed on the poleward side of climatological $\bar{\mathbf{v}}\cdot\bar{T}$ (Fig. 10c). For the eddy momentum flux, the regressed 300-hPa $\bar{\zeta} u'$ exhibits negative anomalies mainly confined to the north of 50°S, which thus lead to the strengthened convergence of eddy westerly momentum (Fig. 10d). Moreover, the anomalous $\mathbf{E}$ vectors in the SH show noticeable equatorward propagation along the entire latitudinal circle north of 55°S, which indicates that wave energy diverges away from the storm-track regions. The divergence of $\mathbf{E}$ vectors corresponds to enhanced westerly winds, consistent with the convergence of eddy westerly momentum in this region. Therefore, the above results demonstrate that SSTID variability is accompanied by distinct variations in storm-track activity, the corresponding eddy fluxes, and wave energy propagation over the two hemispheres.

From an energetics viewpoint, the maintenance of storm tracks requires a continuous extraction of energy from planetary-scale mean flows, which act as baroclinic sources to fuel the development of transient eddies. Thus, looking further into local baroclinic energy conversions over storm-track regions can provide another perspective on SSTID-induced storm-track variations. The release of available potential energy over the extratropics involves conversions from MAPE to EAPE and then from EAPE to EKE. The climatological patterns of baroclinic energy conversions are depicted in contours of Fig. 11. The strongest baroclinic energy conversions occur primarily over storm-track regions in both hemispheres, providing major energy sources for the development of baroclinic eddies that constitute the storm tracks. As a consequence, any fluctuation in storm-track activity will be reflected in local baroclinic energy conversions, and thus storm-track variations can be examined from the perspective of atmospheric energy conversions.

As seen in Fig. 11a, the regression map of energy conversions from MAPE to EAPE shows significant negative anomalies in storm-track regions over both the

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**Fig. 10.** As in Fig. 9, but over the SH extratropics.
North Pacific and North Atlantic, which indicates a decrease of EAPE supply for transient eddies during positive phases of the SSTID. Moreover, the regressed field of baroclinic energy conversions from EAPE to EKE in the NH bears a remarkable resemblance to that of conversion from MAPE to EAPE, which means that, accompanying the reduced supply of EAPE from MAPE, the baroclinic energy conversions from EAPE to EKE are correspondingly reduced (Fig. 11b). As such, diagnostic analysis in terms of local baroclinic energy conversions further supports the conclusion that NH storm-track activity weakens during positive phases of the SSTID. To further reveal the changes in SH baroclinic energy conversions associated with the SSTID, Figs. 11c and 11d show regression maps of SH baroclinic energy conversions onto the normalized SSTID index. During positive phases of the SSTID, both the baroclinic conversions from MAPE to EAPE and from EAPE to EKE over SH storm-track regions are significantly enhanced, which thus promotes the generation and development of transient eddies embedded in SH storm tracks, consistent with the analysis of eddy statistics shown in Fig. 10.

d. Changes in midlatitude transient eddy feedback accompanying SSTID variability

The SSTID variability is accompanied by significant changes in storm tracks and corresponding eddy vorticity and heat fluxes, which can induce anomalous time-mean flows through the transient eddy feedback. Next, the anomalous eddy feedback forcing linked to the SSTID is examined by diagnosing the eddy-induced geopotential height tendencies. Figure 12 depicts anomalous tendencies of height induced by eddy vorticity forcing, thermal forcing, and combined eddy vorticity and thermal forcing at both the lower and upper troposphere in the NH. As presented in Figs. 12a and 12b, the anomalous geopotential height tendencies induced by eddy vorticity flux show an equivalent barotropic pattern throughout the troposphere; meanwhile, the vorticity flux forced tendencies are found to be stronger in the upper troposphere. Whereas the height tendencies forced by eddy heat flux display baroclinic responses...
with nearly opposite patterns between the lower and upper troposphere (Figs. 12c,d). In the lower troposphere, the tendency induced by eddy heat flux is similar in pattern to that forced by vorticity flux, so they reinforce each other (Fig. 12e). At the upper troposphere, although the heat flux–forced tendency is opposite to the vorticity flux–forced tendency, the latter is stronger than the former, so net geopotential height tendency largely retains characteristics of vorticity flux–forced pattern (Fig. 12f). Therefore, the anomalous geopotential height tendencies induced by transient eddy forcing in the NH show equivalent barotropic structures with the meridional dipolar pattern over both the North Pacific and North Atlantic regions.

Moreover, the anomalous fields of eddy-induced geopotential height tendencies over the SH are investigated in Fig. 13. At the lower troposphere, the height tendency induced by eddy vorticity flux is strongest over the midlatitudes (Fig. 13a), while the eddy heat flux forced tendency has the largest magnitude over the polar regions (Fig. 13c), so both eddy vorticity and heat fluxes play important roles in forming the positive SAM-like tendency pattern in the lower troposphere (Fig. 13e). At the upper level, the eddy vorticity flux is dominant in forcing the positive SAM-like tendency, and the contribution from eddy heat flux is very limited (Figs. 13b,d,f). Based on above analysis, it is found that the SSTID-related anomalous eddy feedback forcing induces barotropic tendencies of geopotential height over the two hemispheres. Meanwhile, the patterns of anomalous geopotential height tendency are rather similar to that of extratropical circulation responses to SSTID forcing, which indicates the SSTID influences extratropical large-scale circulation by modulating storm tracks and changing the corresponding transient eddy feedback.

5. Summary and discussion

Aside from the global warming signal, there also exists an equatorial asymmetric mode in global SST variability,
which is characterized by opposite-signed SST anomalies between the two hemispheres and referred to as the SSTID. The influences of SSTID-related asymmetric thermal forcing on wintertime extratropical circulation are investigated by both regression analysis and SST-forced AGCM simulations. It is found that atmospheric responses to the SSTID bear an equivalent barotropic structure over the extratropics. In the SH, positive SSTID events lead to a meridional dipolar perturbation in SLP, which is similar in pattern to the positive SAM. For the NH, SSTID-forced SLP anomalies do not exhibit a zonally symmetric pattern as is the case over the SH; however, they still show signs of a meridional dipole opposite to the SH over the oceans. To understand the dynamical mechanisms responsible for divergent circulation responses to SSTID forcing over the two hemispheres, SSTID-related changes in oceanic fronts, storm tracks, and the transient eddy feedback are examined. It is suggested that the SSTID mode leads to different storm-track variations over the two hemispheres by impacting the intensity of oceanic fronts, which thus induce divergent responses in time-mean

![Diagram](image-url)
flows through the transient eddy feedback. Figure 14 gives a schematic that summarizes the modulating effects of the SSTID on extratropical atmospheric variability. During positive phases of the SSTID, nonuniform SST warming in the NH decreases oceanic fronts in both the North Atlantic and North Pacific basins, and thus leads to the weakening of storm tracks by decreasing atmospheric baroclinicity. The decreased meridional eddy heat flux and momentum convergence then force a north–south dipolar perturbation in geopotential height with positive anomalies to the north and negative anomalies to the south, which thus corresponds to the equatorward (poleward) shift of midlatitude jet over the North Atlantic (Pacific) region. For the SH, the strongest SST cooling in the Southern Ocean causes an intensification of oceanic fronts and atmospheric baroclinicity. The anomalous eddy heat and momentum fluxes accompanying the enhanced SH storm tracks then induce a positive SAM-like pattern in geopotential height, which leads to the poleward shift of westerly jet in the SH.

On the climate impacts of the SSTID, previous studies primarily focused on the tropics and found the SSTID mode can induce equatorial asymmetric responses in tropical rainfall and circulation (Chiang and Friedman 2012; Guo et al. 2016; Sun et al. 2013; Lopez et al. 2016). However, potential influences of the SSTID on extratropical atmosphere remain unclear. In present study, we demonstrated that SSTID-related asymmetric thermal forcing can exert divergent impacts on extratropical circulation over the two hemispheres. The findings in present study further extend our knowledge of the SSTID’s climate impacts. The SSTID mode can simultaneously modulate extratropical circulations over the two hemispheres, and therefore the SSTID may provide dynamical linkage for extratropical circulation variability over the two hemispheres. Moreover, the results suggest that SSTID-induced changes in midlatitude SST gradients play an important role in modulating extratropical circulations, which indicates that we should care more about the distribution of SST rather than SST itself when investigating atmospheric responses to extratropical SST anomalies.

In present study, we primarily investigated one-way influences of the SSTID mode on extratropical circulation; however, anomalous surface winds induced by the SST forcing may also introduce feedback effects.
Affected by the SSTID, surface zonal winds decrease over the subpolar North Atlantic (Fig. 5b). The reduced surface wind speed further strengthens positive subpolar SST anomalies and weakens meridional SST gradients, thus introducing a positive feedback. Over the North Pacific sector, however, the SSTID-induced changes in surface winds are weak and insignificant (Fig. 5c) and therefore the atmospheric feedback effects may be not obvious. For the SH, positive SAM-like responses correspond to increased surface winds over the latitudinal band of 50°–60°S (Fig. 5a), which further cool negative SST anomalies in SH subpolar basin and also introduce a positive feedback. Such positive feedback between the SAM and subpolar SST anomalies in the SH has been documented in a previous study (Xiao et al. 2016). The positive feedbacks between the SST and extratropical atmospheric circulation may help amplify atmospheric responses to SSTID forcing, and this may explain why circulation responses to SSTID forcing in AGCM simulations are weaker than in observations.

Different from that in the SH, extratropical circulation responses to the SSTID in the NH exhibit more regional features; this difference may be associated with distinct land–sea distribution in the two hemispheres. Moreover, it is noted that air temperature responses show subpolar cooling over the North Pacific, which is opposite to that over the North Atlantic region (Figs. 8a,c). For the North Pacific sector, regions north of 65°N covered by sea ice in boreal winter (Fig. 15a), so the subpolar cooling over the North Pacific may be related to variations in Arctic sea ice. Figure 15b shows the correlation map between the SSTID index and sea ice concentration in the NH. In association with the SSTID, sea ice shows contrasting changes between the North Pacific and North Atlantic. Because increased (decreased) sea ice reflects more (less) solar radiation and thus has cooling (warming) effects on atmospheres, distinct sea ice variations over the two basins can be responsible for opposite air temperature responses to SSTID. In the North Atlantic, the SSTID can directly

![Figure 15](image_url)
influence sea ice concentration through SST anomalies, while impacts of the SSTID on Arctic sea ice over the North Pacific sector are very limited due to the barrier effect of land, so changes in sea ice over the North Pacific are mainly controlled by own polar processes. During the twentieth century, there exists a bipolar seesaw of Arctic and Antarctic surface air temperatures, when the Antarctic warms the Arctic cools and vice versa (Chylek et al. 2010). In this study, we found that the SSTID has opposite impacts on polar air temperature between the NA and SH, so the SSTID may help explain the formation of the air temperature bipolar seesaw. Relations between the SSTID mode and polar climate variability deserve further investigation. Because the SSTID includes decadal variability, the limited time span of observational data may bring some uncertainties to our analysis. Especially for the SH, because of sparse station observation, reanalysis data are less reliable before the satellite era. Therefore, the robustness of our results need to be further confirmed with longer reliable observations in the future.

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