Significance of a Midlatitude SST Frontal Zone in the Formation of a Storm Track and an Eddy-Driven Westerly Jet*

TAKEAKI SAMPE
International Pacific Research Center, SOEST, University of Hawaii at Manoa, Honolulu, Hawaii

HISASHI NAKAMURA+
Department of Earth and Planetary Science, University of Tokyo, Tokyo, Japan

ATSUSHI GOTO
Climate Prediction Division, Japan Meteorological Agency, Tokyo, Japan

WATARU OHFUCHI
Earth Simulator Center, Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan

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ABSTRACT

In a set of idealized “aquaplanet” experiments with an atmospheric general circulation model to which zonally uniform sea surface temperature (SST) is prescribed globally as the lower boundary condition, an assessment is made of the potential influence of the frontal SST gradient upon the formation of a storm track and an eddy-driven midlatitude polar front jet (PFJ), and on its robustness against changes in the intensity of a subtropical jet (STJ). In experiments with the frontal midlatitude SST gradient as that observed in the southwestern Indian Ocean, transient eddy activity in each of the winter and summer hemispheres is organized into a deep storm track along the SST front with an enhanced low-level baroclinic growth of eddies. In the winter hemisphere, another storm track forms just below the intense STJ core, but it is confined to the upper troposphere with no significant baroclinic eddy growth underneath. The near-surface westerlies are strongest near the midlatitude SST front as observed, consistent with westerly momentum transport associated with baroclinic eddy growth underneath. The near-surface westerlies are strongest near the midlatitude SST front as observed, consistent with westerly momentum transport associated with baroclinic eddy growth underneath. The sharp poleward decline in the surface sensible heat flux across the SST frontal zone sustains strong near-surface baroclinicity against the relaxing effect by vigorous poleward eddy heat transport. Elimination of the midlatitude frontal SST gradient yields marked decreases in the activity of eddies and their transport of angular momentum into midlatitudes, in association with equatorward shifts of the PFJ-associated low-level westerlies and a subtropical high pressure belt, especially in the summer hemisphere. These impacts of the midlatitude frontal SST gradient are found to be robust against modest changes in the STJ intensity as observed in its interannual variability, suggesting the potential importance of midlatitude atmosphere–ocean interaction in shaping the tropospheric general circulation.

1. Introduction

Synoptic-scale transient eddies are one of the essential components of the tropospheric general circulation, because they transport a substantial amount of heat, moisture, and angular momentum meridionally. The momentum transport maintains midlatitude westerlies, which can be organized as an eddy-driven jet, called a “polar front jet” (“PFJ”; also called a subpolar jet; Palmén and Newton 1969). The eddy activity reaches a maximum

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+ Additional affiliation: Research Institute for Global Change, JAMSTEC, Yokohama, Japan.

Corresponding author address: Takeaki Sampe, Research Center for Advanced Information Science and Technology, University of Aizu, Tsuruga, Ikkima-chi, Aizu-wakamatsu City, 965-8580 Japan. E-mail: sampe@u-aizu.ac.jp

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over the midlatitude oceans where it forms “storm tracks” (Blackmon et al. 1977; Trenberth 1991). In linear theories of baroclinic instability of the zonally uniform westerlies (Charney 1947; Eady 1949), the growth rate of the most unstable normal mode is proportional to the vertical shear of the westerly winds (or, equivalently, the meridional temperature gradient). Therefore, one may consider that an intense subtropical jet (STJ) in the winter hemisphere is particularly favorable for the growth of baroclinic eddies because of the strong vertical shear below its core. In fact, in their numerical experiments Lee and Kim (2003, hereafter LK03) found that baroclinic eddy growth and the associated formation of a PFJ are controlled primarily by the strength and latitudinal position of a STJ. In their experiments, the simulated eddy growth was most enhanced below the intense STJ core, and thus no well-defined PFJ formed in midlatitudes. They found that both a storm track and PFJ develop in the midlatitudes only when the STJ is weak.

In the real atmosphere, however, an intense STJ does not necessarily favor baroclinic eddy growth. For example, baroclinic growth of transient eddies in the wintertime Southern Hemisphere (SH) is weak along an intensified STJ, which merely acts as an excellent waveguide for upper-level transient eddies (Nakamura and Shimpo 2004, hereafter NS04). Rather, NS04 and Nakamura et al. (2004) found that the core region of the SH storm track and PFJ coincides with the Antarctic polar frontal zone (APFZ), which is a prominent oceanic frontal zone in the Indian Ocean extending almost zonally near 50°S along the Antarctic Circumpolar Current (ACC; Colling 2001). NS04 found that the poleward eddy heat fluxes in the SH storm-track core are highly correlated with the local near-surface baroclinicity in their climatological seasonal march. Their finding suggests that vigorous eddy growth in the SH storm-track core is due to the enhanced near-surface baroclinicity associated with the sharp sea surface temperature (SST) gradient across the APFZ, which has been confirmed by atmospheric general circulation model (AGCM) experiments by Inatsu and Hoskins (2004).

In addition, Nakamura and Sampe (2002, hereafter NS02) pointed out that an excessively strong STJ in the midwinter North Pacific hinders baroclinic eddy growth despite the associated strong baroclinicity. They showed that transient eddies are likely to be trapped into the jet core, away from the surface baroclinic zone around 40°N that is anchored by a subarctic oceanic frontal zone (SAFZ; Yasuda et al. 1996; Yuan and Talley 1996). Numerical experiments by Inatsu et al. (2003), Inatsu and Hoskins (2004), and Brayshaw et al. (2008) suggested the significance of zonal variations in the midlatitude SST for the localization of a lower-tropospheric storm track. Furthermore, climate simulations over the North Atlantic for the Last Glacial Maximum and present day (e.g., Kageyama et al. 1999) also indicated the correspondence between sharp surface temperature gradient and a storm track with a westerly jet aloft. These results suggest that, unlike in the idealized experiments by LK03, storm tracks do not necessarily develop preferentially under an intense STJ. Rather, as indicated by observational studies (Sinclair 1995; NS04; Nakamura et al. 2004), core regions of lower-tropospheric major storm tracks tend to coincide with oceanic frontal zones with sharp SST gradients.

Meanwhile, a storm track can influence the ocean by modulating the surface heat and momentum exchanges (Inatsu et al. 2003) through the following processes. Westerly angular momentum transport through horizontal dispersion of baroclinic waves maintains the upper-level westerlies along midlatitude storm tracks, while acting to decelerate STJs (Hartmann 2007) that are maintained by angular momentum transport by the Hadley circulation (Held and Hou 1980). Poleward heat transport associated with baroclinic eddies acts to relax meridional temperature gradients and the vertical shear of the midlatitude westerlies equivalently, leading to the downward transport of the mean westerly momentum to the surface (Lau and Holopainen 1984). The surface westerlies maintained by the momentum transport drive oceanic currents. Based on the interaction of storm tracks and the underlying ocean, self-maintenance of the storm tracks has been discussed in the literature. Hoskins and Valdes (1990) argued that the abundant moisture supply from the oceanic western boundary currents is important for strong diabatic heating in storm-track regions, which excites planetary waves that enhance baroclinicity around the storm-track regions. Nakamura et al. (2004, 2008) argued that, because of the large heat capacity of the oceanic mixed layer and the strong thermal advection by confluent ocean currents, the presence of an oceanic front can exert a strong restoring effect on baroclinicity of the overlying atmosphere that is subject to relaxing by poleward eddy heat fluxes, regardless of the presence of planetary waves.

As is well known, the surface westerlies are of great importance in the general circulation of the ocean. For example, they play a major role in driving the ACC (Gnanadesikan and Hallberg 2000) and its changes (e.g., Oke and England 2004), which might influence the storage of heat and greenhouse gases in the Southern Ocean (Russell et al. 2006). Therefore, understanding the driving mechanisms of the midlatitude surface westerlies is of certain significance for interpreting the simulated climate changes in coupled GCMs.
Although the possible importance of midlatitude SST distribution in locating a storm track has been recently suggested in observational studies (e.g., NS04; Nakamura et al. 2004), the interactive nature of a storm track and the background flow including planetary waves makes it difficult to assess the importance of a midlatitude oceanic front on the formation of storm tracks and westerly jets. In this study, we investigate an impact of the midlatitude SST gradient that could be exerted on the formation of a storm track and midlatitude PFJ, and the robustness of that impact against fluctuations in a STJ. We believe that the impact of a midlatitude SST gradient can be extracted most clearly in the comparison between AGCM experiments with and without frontal gradients in zonally uniform SST profiles. Thus, we perform a set of AGCM experiments with an idealized setting with its lower boundary covered by the zonally uniform ocean, as in Brayshaw et al. (2008). Unlike in their experiments or those of LK03, however, realistic meridional SST profiles with frontal gradients in the midlatitudes are prescribed in some of our experiments. We examine how the simulated circulation in a statistically steady state depends on the intensities of the midlatitude SST gradient and a STJ. While a few highlights of our numerical study have been published in Nakamura et al. (2008), the present paper presents detailed comparisons of the SST frontal influence with the influence of the Hadley cell intensity and extensive analyses of the maintenance of the midlatitude westerlies and storm-track activity, including the estimation of the angular momentum transport and the effect of the surface heat fluxes.

2. Experimental design

The AGCM used in this study is called the AGCM for the Earth Simulator (AFES), whose original code was developed jointly by the Center for Climate System Research (CCSR) of the University of Tokyo and the Japanese National Institute for Environmental Studies (NIES). The model code has been optimized to achieve the best computational efficiency on the Earth Simulator (Ohfuchi et al. 2004, 2007). Physical parameterization schemes adopted in the model include Emanuel’s (1991) scheme for cumulus convection, a radiation transfer scheme by Nakajima and Tanaka (1986), and Louis’ (1979) scheme for surface fluxes. To resolve the observed sharp SST gradient confined in narrow frontal zones, we set the model resolution to be the triangular truncation at the total wavenumber of 79, with 48 sigma levels (T79 L48). The overall experimental design was adopted from that for the “aquaplanet experiment” project proposed by Neale and Hoskins (2000), except for the insolation. The insolation in our experiments is fixed to the boreal summer solstice condition. As in Hayashi and Sumi (1986), the zonally uniform SST was prescribed globally at the model’s lower boundary with no landmass.

In our experiments, we attempt to differentiate between the influence of the midlatitude SST gradient and that of a STJ on the extratropical atmosphere by assigning different meridional SST profiles. As shown in Fig. 1, and as summarized in Table 1, we adopted six SST profiles, each of which consists of a combination of one of the two profiles for midlatitudes and one of the three for the tropics. In our control (CTL) experiment, we used a SST profile based on the climatology observed in the southwestern Indian Ocean, where the core of the SH storm track is observed year-round (NS04). The meridional SST profile was taken from the optimum interpolation (OI) SST version 2 for the 1982–95 period, provided by the National Oceanic and Atmospheric Administration (NOAA). In this particular dataset in situ ship observations and satellite measurements have been blended (Reynolds et al. 2002). In the CTL experiment, the 14-yr climatology of June–August SST averaged between 60° and 80°E was given over the
model SH, and the corresponding December–February climatology was given over the model Northern Hemisphere (NH). The SH and NH of the model atmosphere thus correspond to the winter and summer hemispheres, respectively. In this profile SST exhibits a sharp decline at midlatitudes with gradients of 1.8°C (110 km)−1 at 45°S and 1.7°C (110 km)−1 at 45°N. Any observed SST values below 0°C have been replaced with 0°C in this profile in order to realize an ice-free condition.

In our model, the STJ intensity is controlled primarily by thermal contrasts between the tropics and subtropics through modulations in the Hadley cell intensity. In a “strong STJ” (“SJ”) experiment, tropical SST was raised from the CTL profile to yield a well-defined peak at 5°N (Fig. 1), which resulted in the enhancement of STJs in the two hemispheres. Specifically, the SST profile $T_{SJ}$ (°C) in the tropics was given by

$$T_{SJ}(\phi) = 31 \left[ 1 - \sin^2 \left( \frac{\phi - \phi_m}{\phi_p} \right) \right].$$

(1a)

where $\phi$ is latitude and $\phi_m = 5^\circ$. In (1a), $\phi_p = 110^\circ$ is set for $5^\circ \leq \phi \leq 18^\circ$ (NH) and $\phi_p = 130^\circ$ for $-12^\circ \leq \phi \leq 5^\circ$ (mainly SH). In a “weak STJ” (“WJ”) experiment, tropical SST was lowered to weaken the Hadley circulation, especially in the winter hemisphere. The particular SST profile $T_{WJ}$ (°C) in the tropics was given by

$$T_{WJ}(\phi) = 18 + 9 \left[ 1 - \sin^2 \left( \frac{\phi - \phi_m}{\phi_p} \right) \right].$$

(1b)

In (1b), $\phi_p = 110^\circ$ is set for $5^\circ \leq \phi \leq 30^\circ$ (NH) and $\phi_p = 130^\circ$ for $-16^\circ \leq \phi \leq 5^\circ$ (mainly SH).

For another experiment (hereafter “NF” experiment), the midlatitude frontal SST gradient in the CTL profile was replaced by a uniform gradient of 0.3°C (110 km)−1 for the model SH and 0.35°C (110 km)−1 for the model NH (Fig. 1). In another two experiments, SST was raised and lowered from the NF profile only in the tropics, just as in the SJ and WJ experiments (hereafter “NFSJ” and “NFWJ” experiments, respectively). Our NF, NFSJ, and NFWJ profiles are characterized by weaker SST gradients in midlatitudes. In addition, SST poleward of 45°S and 45°N is much warmer relative to the CTL profile, which might have some side effects, including too much moisture supply from the subpolar ocean. Nevertheless, we did not lower the midlatitude SST, because it would lead to the enhancement of subtropical SST gradients or the cooling of tropical SSTs, which would change the STJ structure and/or the Hadley cell intensity.

For each of the six experiments, we analyze the model output sampled every 6 h for a 24-month period from the 7th to the 30th month of the perpetual integration. The transient eddy signal for a given variable is defined as high-frequency fluctuations extracted with a high-pass filter with the half-power cutoff period of 8 days. Each of the zonally averaged eddy statistics presented below to characterize the storm-track activity has been calculated from those fluctuations over the 24-month period.

### 3. Mean state of model storm tracks and westerly jets

#### a. CTL experiment with midlatitude frontal SST gradient

Figure 2 shows the zonally averaged time-mean statistics of transient eddy amplitude, the Eliassen–Palm (E–P) flux, zonal wind speed, and mean-flow baroclinicity in the CTL experiment for the winter (Figs. 2a,c) and summer (Figs. 2b,d) hemispheres. The eddy amplitude is measured as the root-mean-square (rms) of the meridional wind velocity fluctuating with transient eddies over the 24-month period. The mean-flow baroclinicity is defined as the Eady (1949) growth rate $\gamma_{BI} = 0.31/\sqrt{\frac{\partial U/\partial z}{N^2}}$, where $\overrightarrow{U}$ is the zonal-mean zonal wind velocity, $N$ the buoyancy frequency, and the overbar the time averaging. The E–P flux is defined here as $(E_x, E_z) = \rho \partial \mathbf{u} \cdot \mathbf{v} \rho_0 \cos \phi - \mathbf{f} \cdot \overrightarrow{v} \mathbf{v} / \Theta_z$, where $\rho_0$ (1000 hPa) is the standard surface pressure and $\Theta_z$, vertical derivative of potential temperature of the mean state (Andrews and McIntyre 1976; Plumb 1986). The E–P flux represents the propagation of wave activity, and its direction is parallel to the group velocity of the wave. Effective forcing of $\overrightarrow{U}$ by eddies is proportional to $\mathbf{V} \cdot \mathbf{E}$ in the transformed Eulerian-mean framework, as they transport mean-flow westerly momentum in the direction opposite to $\mathbf{E}$.

In the time-mean state over the winter hemisphere (SH in our model), the primary storm track is organized into a deep structure at the midlatitudes (Fig. 2a), and its
axis is located slightly south of the sharp SST gradient (Fig. 2c). A snapshot of the near-surface meridional wind velocity is shown in Fig. 3c as an example of synoptic-scale eddies, which appear to be active around the mid-latitude oceanic frontal zone. Correspondingly, the upward E–P flux (or, equivalently, the poleward eddy sensible heat flux) in the lower troposphere is strongest around 45°S. Because the poleward eddy heat flux represents the conversion of available potential energy (APE) from the mean state into eddies, the large heat flux in mid-latitudes indicates the intense baroclinic growth of transient eddies (Fig. 2a). In addition, another storm track forms at 28°S, with its core situated just below the intense STJ core. The core of this subtropical storm track is at

\[2\] At the surface, the profile of the wind variance exhibits a slight bump (dip) on the equatorward (poleward) flank of the SST front, as observed by Booth et al. (2010), resulting from such boundary layer processes as momentum mixing intensity varying with surface stability (not shown).
the 200-hPa level, which is higher by $\sim 1.5$ km than the midlatitude storm-track core in correspondence to the higher tropopause in the subtropics. This subtropical storm track exhibits a shallow structure with eddy amplitude decaying rapidly downward, as actually observed in the wintertime SH (NS04). Correspondingly, the upward E–P flux is weak in the lower troposphere, indicating weak baroclinic eddy growth in spite of strong mean baroclinicity in the free troposphere below the STJ core (Fig. 2c). The strong upper-level eddy activity in the subtropics is sustained mainly by the influx of wave activity from the midlatitude storm track, as indicated by the converging equatorward E–P flux (Fig. 2a). These characteristics of the subtropical storm track are representative of the trapping effect of wave activity by a strong STJ core, as observed by NS02 for the wintertime North Pacific and by NS04 for the wintertime South Pacific. In fact, the maximum wind speed of the model STJ is greater than 50 m s$^{-1}$ (Fig. 2c), which is similar to that observed for the wintertime South Pacific STJ ($\sim 54$ m s$^{-1}$).

In the mean state for the summer hemisphere (NH in our model), the Hadley cell and STJ are much weaker (Fig. 2d) than in the winter hemisphere, as observed. A single, well-defined storm track forms around 50°N (Fig. 2b), slightly poleward of the latitude of the sharp SST gradient (Fig. 2d). This storm track exhibits a deep structure with the pronounced lower-tropospheric upward
E–P flux that indicates intense baroclinic eddy growth above the midlatitude surface baroclinic zone. A tendency for synoptic-scale eddies to develop around the midlatitude oceanic frontal zone is also seen in a snapshot of the near-surface meridional wind velocity (Fig. 3a). A STJ still exhibits its core at around 30°N, with large baroclinicity underneath, but the jet appears to be too weak to trap wave activity. In fact, no eddy activity maximum is seen around the STJ core (Fig. 2b). It is noteworthy that the maximum amplitude of transient eddies is greater in the summer hemisphere, where the STJ and associated midtropospheric baroclinicity are weaker than in the winter hemisphere. The vigorous baroclinic eddy growth at midlatitudes must be sustained by strong near-surface baroclinicity above the sharp SST front around 45°N and associated APE.

In the CTL experiment, the axis of the low-level westerlies is not located below the STJ core in either hemisphere. In the summer hemisphere, a PFJ is the dominant jet located around 50°N, which is maintained through the westerly momentum transport from the subtropics associated with the equatorward propagation of baroclinic wave activity, as indicated by the equatorward E–P flux emanating from the midlatitude storm track (Fig. 2b; see the definition of the flux in the first paragraph of this section). The PFJ exhibits a deep structure with the well-defined axis of the low-level westerlies under its upper-level jet core. In the winter hemisphere, an upper-level PFJ is seen as a secondary peak in the westerly wind speed, though it is not completely separated from the dominant STJ. In the lower troposphere, a westerly jet is located around 48°S, in correspondence with the midlatitude storm track and the oceanic frontal zone. It is therefore suggested that the downward transport of westerly momentum by storm-track activity maintains the low-level westerly jet at midlatitudes. It should be noted that the mean surface westerly wind almost vanishes under the STJ, indicating that the axis of the subtropical high pressure belt is situated below the STJ core.

A comparison of Fig. 2 with observed circulation fields over the Indian Ocean (Fig. 3 of NS04) reveals a certain resemblance between the modeled and observed structure of the midlatitude storm track and westerlies in the summer hemisphere. In the winter hemisphere, the simulated westerly wind field differs somewhat from that observed, probably due to our prescription of the observed Indian Ocean SST on all longitudes in the model. The simulated wintertime STJ is so strong that the secondary upper-level storm track forms in the subtropics, but otherwise the midlatitude circulation is similar to the observed field.

b. Effects of the elimination of the midlatitude frontal SST gradient (NF experiment)

Figure 4 shows the zonal-mean statistics of transient eddies and the mean flow simulated in the NF experiment in the same manner as in Fig. 2. A comparison between the two figures reveals profound changes in eddy activity and the midlatitude westerlies as the response to the removal of the frontal SST gradient. In the winter hemisphere of the NF experiment, the midlatitude storm-track core marked by a local upper-level maximum in eddy activity becomes less distinct while shifted equatorward, in association with ~25% reduction in eddy amplitude relative to the CTL experiment (Fig. 4a). In fact, a snapshot of the low-level meridional wind velocity shown in Fig. 3d indicates weaker and less organized eddy activity. The reduction in eddy amplitude leads to marked weakening of the poleward eddy heat transport and the associated upward E–P flux in midlatitudes (Figs. 2 and 4). Figure 5 shows latitudinal profiles of transient eddy activity in the upper and lower troposphere, measured as the rms of the meridional wind velocity and the poleward eddy heat flux, respectively, for our six experiments. The peak in the eddy heat flux in the NF experiment is much less distinct, with the maximum value ~70% smaller than that in the CTL experiment (Fig. 5c), resulting from a reduction of APE and its conversion associated with the smoothed SST gradient in the midlatitudes. A comparison between the CTL and NF experiments indicates that the sharp SST gradient associated with a midlatitude oceanic frontal zone favors baroclinic eddy growth, leading to the organization of transient eddy activity into a prominent midlatitude storm track. In the subtropics, the difference in the eddy amplitude between the two experiments is rather small.

In the NF experiment the elimination of the midlatitude frontal SST gradient also exerts a pronounced impact on the westerly jets in the winter hemisphere, including the diminishing of a PFJ and the equatorward shift of the low-level westerly belt (Fig. 4c). Meridional profiles of the mean zonal wind speeds at the 200- and 925-hPa levels are plotted in Fig. 6. Relative to the CTL experiment, the midlatitude westerlies in the NF experiment are weakened by ~10 m s⁻¹ in the upper troposphere and by 5–8 m s⁻¹ in the lower troposphere, but are intensified in the subtropics (Figs. 6a,c). These changes are consistent with the weakening of midlatitude eddies that transport westerly momentum from the subtropics into the midlatitudes. In fact, the equatorward

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3 In reality, the secondary upper-level storm track along the STJ is observed only in the Australian–Pacific sector (NS04).
component of the E–P flux ($E_v$) in the upper troposphere is weaker than in the CTL experiment (Fig. 4a), and the reduced westerly acceleration in the midlatitude lower troposphere as indicated by the weakened upward E–P flux (Figs. 4a and 5c) is consistent with the weaker near-surface westerly wind. Correspondingly, the axis of the near-surface westerlies is shifted equatorward by $\sim 10^\circ$. This shift is not attributable to the reduction of vertical shear of the westerlies between the 925-hPa level and the surface resulting from the removal of the sharp temperature gradient, because the 1000-hPa westerly axis is also shifted equatorward. The change in the low-level westerlies is accompanied by an equatorward axial shift (by $\sim 3^\circ$) of the subtropical high pressure belt, marked as the zero-mean zonal wind velocity in Fig. 6c.

Differences between the CTL and NF experiments are also evident in the summer hemisphere. In the upper troposphere, the elimination of the midlatitude frontal SST gradient leads to a substantial decrease in eddy amplitude at the mid- and high latitudes with a storm-track core weakened by $\sim 30\%$ and shifted equatorward by $\sim 10^\circ$ (Figs. 4b and 5b). In the NF experiment, low-level eddy activity measured by poleward eddy heat flux is also weakened markedly at midlatitudes with its peak magnitude reduced by as much as $\sim 70\%$ (Fig. 5d). The weakening of eddy activity and scattering in the latitudes of the individual eddy centers are manifested in

FIG. 4. As in Fig. 2, but for the NF experiment.
the snapshot of eddies at a near-surface level (Fig. 3b). Correspondingly (Figs. 4d and 6b,d), the axes of the upper-level jet and low-level westerlies are both shifted equatorward by $10^\circ$, and so is the axis of the subtropical high pressure belt ($7^\circ$ in latitude). Again, the shift of the low-level westerly axis is not attributable to the reduced vertical shear of the near-surface westerlies associated with the relaxed SST gradient (Fig. 6d). Overall, the removal of the midlatitude frontal SST gradient results in the weakening of midlatitude storm-track activity and westerly wind and the equatorward shifts of the entire circulation system in the extratropical troposphere. It is noteworthy that eddy activity is weaker in the NF experiment in spite of the stronger midtropospheric baroclinicity in either hemisphere (cf. Figs. 2 and 4), which indicates that midtropospheric or vertically averaged baroclinicity is not as important as the near-surface baroclinicity in maintaining storm-track intensity.

c. Influence of the anomalous STJ intensity (SJ–WJ and NFSJ–NFWJ experiments)

In the preceding subsections, our comparison between the CTL and NF experiments has revealed the substantial influence exerted by the midlatitude frontal SST gradient on the formation of a storm track and PFJ. At the same time, in each of the two experiments, the equatorially asymmetric SST distribution yields distinct differences in the strength of the Hadley cells and associated STJs; they are much stronger in the winter hemisphere, as observed, leading to the formation of a secondary storm track along the STJ. In the real atmosphere, either in winter or summer, intensities of the Hadley cell and STJ also vary interannually in association, for example, with El Niño–Southern Oscillation (ENSO). In this subsection we assess how robust the influence of the frontal SST gradient could be against such modest changes in the

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**FIG. 5.** Meridional profiles for rms of 250-hPa eddy meridional wind fluctuations (m s$^{-1}$) in the (a) winter and (b) summer hemisphere for the six experiments: CTL (black solid line with open circles), SJ (red solid line), WJ (blue-dotted line), NF (green solid line with triangles), NFSJ (orange-dashed line), and NFWJ (purple dot–dashed line). As in (a), but for 850-hPa poleward eddy flux of sensible heat (K m s$^{-1}$) in the (c) winter and (d) summer hemisphere. The large blue triangles indicate the latitude of the SST frontal zone for the CTL, SJ, and WJ experiments.
Hadley cell and STJ as observed in their interannual fluctuations. Our assessment is made separately for the winter and summer hemispheres by comparing additional sets of our experiments, namely, between the SJ and WJ experiments and between the NFSJ and NFWJ experiments, based on Figs. 5 and 6 and Table 2.

Through our assessment only the limited influence of the modest changes in the Hadley cell and STJ intensities is found on the midlatitude storm tracks and associated PFJs (Figs. 5 and 6). In the winter hemisphere at 49°S, for example, the difference in the upper-level eddy amplitude measured as the rms of meridional velocity is only 0.26 m s⁻¹ (0.05 m s⁻¹) between the CTL and WJ (SJ) experiments, which is less than ~6% of the corresponding difference between the CTL and NF experiments (Table 2). The corresponding ratio of the differences in the 850-hPa poleward eddy heat flux is less than 13% in the winter hemisphere and less than 6% in

Table 2. Differences in the quantities shown in Figs. 5 and 6 and the 850-hPa upward E–P flux, between the CTL experiment and each of the SJ, WJ, and NF experiments sampled along the upper-level storm track axis in the CTL experiment. The (left) differences arising from the enhancement and (middle) weakening of the Hadley cell, and (right) the removal of the midlatitude frontal zones are shown.

<table>
<thead>
<tr>
<th></th>
<th>SJ – CTL</th>
<th>WJ – CTL</th>
<th>NF – CTL</th>
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<tbody>
<tr>
<td>Winter (49°S)</td>
<td></td>
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<tr>
<td>(V_{\text{rms}}) (m s⁻¹)</td>
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<td>-0.26</td>
<td>-4.83</td>
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<td>EP,850 (m² s⁻²)</td>
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<td>0.035</td>
<td>-0.098</td>
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<td>(V'U'850) (K m s⁻¹)</td>
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<td>U200 (m s⁻¹)</td>
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<td>-8.99</td>
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<tr>
<td>U925 (m s⁻¹)</td>
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<td>1.42</td>
<td>-5.80</td>
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<tr>
<td>Summer (50°N)</td>
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<td>(V_{\text{rms}}) (m s⁻¹)</td>
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<tr>
<td>U925 (m s⁻¹)</td>
<td>-3.84</td>
<td>-0.53</td>
<td>-11.70</td>
</tr>
</tbody>
</table>
the summer hemisphere (Table 2). The strongest impact from the tropics is found in the upper-tropospheric eddy activity over the summer hemisphere. In the SJ experiment, the significant weakening of the midlatitude westerlies (Fig. 6b) accompanies a significant reduction in the midlatitude eddy activity (Fig. 5b). Still, the eddy activity reduction reaches only 26% of the corresponding difference between the CTL and NF experiments (Table 2). Otherwise, the modest STJ changes simulated among the CTL, SJ, and WJ experiments can exert only an insignificant impact on the midlatitude storm-track activity, and the corresponding impact on the PFJ and associated low-level westerlies are also weaker than that of the frontal SST gradient in the midlatitudes.

In the subtropics, of course, changes in the intensities of the Hadley cell and associated STJ can influence eddy activity substantially. In the winter hemisphere, the weakened Hadley cell in the WJ experiment results in a reduction of the STJ core velocity (blue-dotted line in Fig. 6a) by as much as 18 m s⁻¹ relative to the CTL experiment to reach the value (~40 m s⁻¹) close to the observation over the southeastern Indian Ocean (NS04). Correspondingly, the secondary storm track is significantly diminished (Fig. 5a). In contrast, the enhanced Hadley cell in the SJ experiment leads to an increase of 11 m s⁻¹ in the STJ core velocity (red line in Fig. 6a) relative to the CTL experiment, but the associated changes in the eddy amplitude are insignificant (Fig. 5a). In the summer hemisphere, the corresponding influence on the upper-level westerlies and eddy activity in the subtropics is stronger in the SJ experiment than in the WJ experiment (Figs. 5b and 6b).

The modifications in the tropical SST in the absence of the frontal midlatitude SST gradient result in an increase (decrease) of the STJ core velocity by ~10 m s⁻¹ in the NFSJ (NFW) experiment relative to the NF experiment in the winter hemisphere (Fig. 6a). However, very little change is found in the storm-track activity and zonal wind velocity in the midlatitudes. The corresponding changes are also small in the midlatitude summer hemisphere, except for a modest increase (~1.5 m s⁻¹) in the upper-level eddy amplitude in the NFSJ experiment (Fig. 5b), again indicating the robustness of the influence of the midlatitude SST gradient.

Over the subtropical southwestern Indian Ocean, the standard deviation of interannual fluctuations observed in upper-level westerly wind speed is at most ~5.5 m s⁻¹ for austral winter and ~4 m s⁻¹ for summer (not shown). Because these changes are smaller than the STJ changes in our AGCM simulations, one may expect fluctuations that would be observed in the midlatitude storm-track activity to be as modest as the differences simulated in our experiments (CTL – SJ – WJ). In fact, the correlation coefficient between the eddy amplitude in high-frequency (subweekly) 250-hPa height fluctuations in the midlatitudes (55°S–45°S) and the 250-hPa westerly wind speed at 27.5°S averaged between 60° and 80°E, based on the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis from 1979 to 1998, is ~0.436 and ~0.234 for austral winter and summer, respectively. This means the observed interannual fluctuations in the STJ intensity can explain only about 19% and 6% of the interannual variances for the respective seasons in midlatitude eddy amplitude in the SH storm-track core over the southwestern Indian Ocean.

**d. Influence of the frontal SST gradient on the global angular momentum budget**

The above-mentioned influence of the midlatitude frontal SST gradient on the storm-track activity and zonal-mean wind is also manifested as significant changes in the global angular momentum budget. Figure 7 shows the mean distributions of the relative angular momentum at the 200-hPa level, the eastward torque by zonal wind acting on the earth surface, and the vertically averaged transport of angular momentum by eddies. In the NF experiment, where eddy activity is unrealistically weak due to the relaxed midlatitude SST gradient, relative angular momentum is concentrated in the subtropics resulting from the transport by the Hadley circulation (Fig. 7a). In the CTL experiment, the marked strengthening of eddies results in the enhanced redistribution of angular momentum from the subtropics into higher latitudes (Fig. 7c), and the consequent formation of midlatitude PFJ (Fig. 7a). At the surface (Fig. 7b), the influence of the SST frontal zone is manifested as a poleward shift of the entire circulation system, including the midlatitude westerlies, the tropical trades and the subtropical high pressure belt in between, and the weakening of the polar easterlies. Another manifestation is the enhanced exchange of angular momentum between the atmosphere and ocean/solid earth that is not only under the poleward-shifted midlatitude westerlies, but is also under the enhanced and widened trades (Fig. 7b), resulting in stronger driving of the ocean circulation in the presence of the frontal SST gradient.

**4. Maintenance of near-surface baroclinicity**

In each of the CTL, SJ, and WJ experiments, the tight SST gradient prescribed in midlatitudes maintains...
near-surface baroclinicity of the model atmosphere through surface sensible heat flux (SHF), whose mean meridional profiles are shown in Figs. 8a,b for the winter and summer hemispheres, respectively. The presence of the frontal SST gradient exerts a profound influence on the surface turbulent heat fluxes. In any of the CTL, SJ, and WJ experiments, the surface SHF peaks at the equatorward flank of the frontal zone, decreasing sharply poleward across it by \( \sim 35 \) and \( \sim 20 \) W m\(^{-2}\) in the winter and summer hemispheres, respectively. Figures 8c,d show the mean difference between zonally averaged SST and surface air temperature (SAT). The latter has been estimated as

\[
\text{SAT} = \sigma_1^{-R/C} T_1,
\]

where \( T_1 \) denotes air temperature at the lowest sigma level (\( \sigma_1 \)) of the model. The mean air–sea temperature difference (positive if SST > SAT) is maximized at the equatorward flank (\( \sim 42^\circ \) latitude) of the frontal zones, corresponding well to the peak of the surface SHF. At the poleward flank (\( \sim 48^\circ \) latitude), in contrast, SAT is higher than SST (Figs. 8c,d), probably resulting from the effect of atmospheric heat transport, and the model atmosphere is cooled by low SST (Figs. 8a,b; cf. Fig. 1), as

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**FIG. 7.** Meridional profiles of (a) the 200-hPa relative angular momentum \([U_{200} \cos(\text{lat})] \quad \text{m s}^{-1}\), (b) the surface westerly torque \([\tau \cos(\text{lat})] \quad \text{N m}^{-2}\), and (c) the vertically averaged eddy transport of angular momentum \([u'v' \cos(\text{lat})] \quad \text{m}^2 \text{s}^{-2}\). Based on the zonally averaged statistics for the CTL (solid line) and NF (dashed line) experiments. Abscissa is scaled proportional to \( \cos(\text{lat}) \).
in the NCEP/Department of Energy (DOE) reanalysis (Kanamitsu et al. 2002; Figs. 8g,h). While the surface turbulent heat fluxes depend on surface wind speed, the distinct similarity between the profiles of the surface SHF and air–sea temperature difference clearly shows the crucial importance of the latter for air–sea heat exchange. The sharp decline in surface SHF across the SST front is consistent with the slightly smoother SAT profile whose meridional gradient across the SST front is 15%–25% weaker than that of the SST profile (Figs. 8e,f; shown only for the CTL and NF experiments), resulting from atmospheric heat transport. Nevertheless, the SAT profile for each of the CTL, SJ, and WJ experiments is characterized by a sharp decline across the midlatitude SST front, which is absent in the experiments without the SST front (i.e., the NF, NJSF, and NFWJ experiments). The frontal SST gradient and the associated sharp decline in heat supply from the ocean thus maintain near-surface baroclinicity. In each of our experiments with no frontal SST gradient, the meridional SAT gradient is only ~25% of that in the CTL experiment at 45°S and ~30% at 45°N (Figs. 8e,f), which is indicative of much smaller near-surface baroclinicity.

An analysis of the effect of eddy heat transport and the surface SHF on the heat budget in the planetary boundary layer (PBL) confirms the importance of the surface SHF in maintaining the near-surface baroclinicity. Figures 9a,b show the direct thermal effect of transient eddies on the meridional temperature gradient averaged within the PBL between the 1000- and 850-hPa levels, measured as the equatorward gradient of the convergence of the eddy sensible heat flux $[-s\partial^2\psi/\partial y^2(T' T')]$, for the model winter (southern) hemisphere (where $s = 1$) and summer (northern) hemisphere (where $s = -1$), respectively. Figures 9c,d present the corresponding effect of the surface SHF $[g/(C_p \Delta p) \times \partial \psi/\partial y$ (SHF); $\Delta p = 150$ hPa] under the assumption that locally it either warms or cools the atmosphere only below the 850-hPa level. In the CTL experiment, transient eddies act to decrease the temperature gradient around the SST frontal zone, mostly due to their poleward heat transport. This relaxing effect by eddies can be compensated by the sharp contrast of the surface SHF across the oceanic frontal zone if the sensible heating is confined to the PBL (Figs. 9c,d; thick line with circles). As shown in Figs. 9c,d, the thermal effect of the secondary circulation through the meridional and vertical temperature advection $\{x\partial \psi/\partial y\} - \nu T' \partial /\partial y - \omega \partial \psi/\partial p (p/p_c)_{\text{RCs}}\}$, evaluated from the low-frequency component (including the time mean) of winds and temperature, does not contribute to the restoration of the near-surface baroclinicity associated with the SST frontal zone. Rather, it counteracts the restoring effect of the surface SHF. Though playing a certain role, possibly in the self-maintenance of storm tracks in the free troposphere (Robinson 2006), the secondary circulation does not contribute to the anchoring of storm tracks above the SST front by maintaining the near-surface baroclinicity. In the NF experiment, both the eddy heat transport and the surface SHF are much more uniform meridionally than in the CTL experiment. Correspondingly, their effects on the meridional temperature gradient within the PBL are much weaker (Fig. 9), consistent with much weaker near-surface baroclinicity (Figs. 8e,f). Although condensation heating can influence the temperature gradient in the PBL, the latent heating is not analyzed here due to a problem of data availability. Shallow cumulus convection can contribute to the restoration of the near-surface baroclinicity if it occurs mainly on the warmer side of the SST frontal zone under the cold-air outbreaks behind the cold front of individual cyclones. In contrast, the formation of stratus on the cooler side of the SST front resulting from advected warm, moist air can act to relax the baroclinicity.

The opposing roles of eddy heat transport and surface SHF in maintaining the SAT gradient can be elucidated by examining their temporal characteristics. Figure 10 shows cross-correlation coefficients between the low-level poleward eddy heat transport and the difference across the SST frontal zone in SAT or surface SHF, calculated from 6-hourly zonal-mean time series separately for the CTL and NF experiments. In either hemisphere, negative correlation of the heat transport with the cross-frontal SAT difference becomes strongest when the latter lags the former by a half-day (Figs. 10a,b). For nearly the same lag, the corresponding positive correlation with the cross-frontal difference in the surface SHF is maximized (Figs. 10c,d). These lead–lag relationships indicate that as baroclinic eddies amplify, their poleward heat transport relaxes the SAT gradient within a day, but the cross-frontal differential heat supply from the ocean builds up simultaneously to restore the SAT gradient. Those positive and negative correlations diminish when the lag reaches 2 or 3 days, suggesting that the heat supply from the ocean could restore the strong cross-frontal SAT gradient efficiently. The timing of this particular diminution lags by 12–24 h behind the diminution of the autocorrelation of the poleward heat transport (not shown), in agreement with the restoring time scale of 1 day estimated by Nonaka et al. (2009) based on a high-resolution coupled GCM simulation. The correlations in the NF experiment are much weaker, which are indicative of weaker restoration of the SAT gradient. Although the surface SHF acts to damp temperature anomalies associated with individual cyclone systems as discussed in Hoskins and Valdes (1990), our results in the preceding section indicate that the differential heat
supply from the ocean across a midlatitude oceanic frontal zone, as a whole, enhances storm-track activity through maintaining a surface baroclinic zone. Recently, Hotta and Nakamura (2009, manuscript submitted to J. Climate) examined the relative importance of the surface SHF in maintaining near-surface baroclinicity among all kinds of diabatic heating in the NCEP/DOE reanalysis through numerical experiments, as in Hoskins and Valdes (1990). Their results suggest that the intense near-surface baroclinicity in the core regions of the major storm tracks cannot be maintained without sensible heat exchanges with the ocean.

Latitudinal distributions of the surface latent heat flux (LHF) for our experiments are compared in Figs. 11a,b. In each of the CTL, SJ, and WJ experiments the LHF decreases sharply with latitude across the frontal zones by as much as 100 W m\(^{-2}\) in the winter hemisphere and \(
\approx 60 \text{ W m}^{-2}\) in the summer hemisphere. An abundant moisture supply into the atmosphere at the equatorward flanks of the frontal zones favors baroclinic eddy growth through enhancing latent heat release.\(^5\) Precipitation rate in the winter hemisphere exhibits a distinct peak at the equatorward flank of the frontal zone, collocated with the peak of the surface LHF (Fig. 11c). There is no signature of the subtropical storm track in the precipitation profiles, consistent with the suppressed baroclinic eddy growth along the STJ. Precipitation rate in the summer hemisphere displays a broad peak around the mid-latitude storm track (Fig. 11d).

\(^5\) As Hoskins and Valdes (1990) suggested, latent heat release along a storm track may act to maintain the mid- to lower-tropospheric baroclinicity by inducing a planetary wave response, but this effect cannot be operative in such a zonally symmetric circumstance as in our experiments.
In the absence of the frontal SST gradient, the surface LHF in each of the NF, NFSJ, and NFWJ experiments decreases more gradually with latitude, with the stronger surface LHF at high latitudes resulting from higher SST than in the CTL, SJ, and WJ experiments. Probably due to the enhanced moisture supply from the warmer ocean surface, no significant decrease in precipitation is simulated in the summer hemisphere despite the substantial weakening of the storm-track activity (Fig. 11d) as the response to the elimination of the frontal SST gradient. In the winter hemisphere, in contrast, the elimination yields a reduction in precipitation at its peak (Fig. 11c), in correspondence with the reduced moisture supply and storm-track activity.

5. Summary and discussion

Though highly idealized, the “aquaplanet” experiments conducted in the present study have demonstrated the potential importance of the midlatitude frontal SST gradient in the formation of storm tracks and westerly PFJs. The strong near-surface baroclinicity associated with the SST front enhances baroclinic eddy growth, leading to the organization of a prominent midlatitude storm track throughout the depth of the troposphere. The storm-track formation results in a westerly acceleration in the midlatitudes through poleward angular momentum transport associated mainly with equatorward wave propagation. The transferred westerly momentum is organized into a well-defined PFJ with a deep structure, as actually observed in the SH (cf. NS04). Therefore, the main storm track, as well as the PFJ, tends to be anchored around the midlatitude SST frontal zone, as postulated by Nakamura et al. (2004). The elimination of the frontal SST gradient in our model yields a substantial reduction in midlatitude eddy activity and the equatorward shift of the entire low-level circulation system, including a storm track, near-surface westerlies, and a subtropical high pressure belt. The impacts of the midlatitude frontal SST gradient simulated in our model are found to be robust against modest changes in the strength of a STJ as observed in its interannual variability over the south Indian Ocean.

Distinct differences in a storm track and PFJ between the winter and summer hemispheres, as observed by NS04 and simulated in the present study, may be influenced by pronounced seasonal differences in the strength of the Hadley circulation and associated STJ. Nevertheless, our results suggest that the STJ strength is not a sole

![Fig. 8. (Continued)](image-url)
factor controlling the location and activity of a storm track and the formation of a PFJ, in agreement with numerical experiments by Brayshaw et al. (2008). In our experiments, the baroclinicity associated with a STJ does not correspond well to actual baroclinic eddy growth and storm-track activity. Specifically, 1) the eddy amplitude is greater in the summer hemisphere where the STJ is less intense, 2) baroclinic eddy growth is inactive below the STJ in both hemispheres, and 3) modest fluctuations in the STJ intensity as observed in its interannual variability exert no substantial influence on the storm-track activity.

Our experiments with a realistic SST profile are in agreement with recent observations (NS02; NS04) that showed the tendency for a STJ to be unfavorable for baroclinic growth of eddies. This tendency is in contrast with the experiments by LK03, where the STJ intensity controls baroclinicity down to the surface and thereby baroclinic eddy growth. In their experiments, the initial surface temperature profile is characterized by a broad baroclinic zone spanning the entire midlatitudes, as in our NF experiment. Thus, there is no strong external forcing that anchors a storm track in the midlatitudes because of the lack of frontal features in surface temperature. In contrast, we prescribed a realistic SST profile with a sharp SST gradient for our CTL experiment, which maintains a surface baroclinic zone along the SST frontal zone. As a result, both a storm track and PFJ are well defined, even in the presence of the enhanced Hadley cell in the winter hemisphere, as observed in the south Indian Ocean (NS04; Nakamura et al. 2004). Thus, our experiments have revealed the particular importance of
midlatitude oceanic frontal zones in the formation of storm tracks and associated PFJs, which has been overlooked in most of the previous modeling studies.

Recently, Son and Lee (2005) suggested that the weak Hadley circulation and broad extratropical baroclinic zone are favorable for the separation of a STJ and PFJ. Our experiments indicate, however, that the SST gradient concentrated in a narrow midlatitude frontal zone, which cannot be represented by the smooth temperature profiles employed in LK03 and Son and Lee (2005), is also of critical importance for the STJ–PFJ separation. We therefore argue that the manner in which near-surface baroclinicity is controlled in an atmospheric model can exert a significant impact on a simulation of storm tracks and midlatitude westerlies. Nevertheless, the impact of the frontal SST gradient simulated in the present study should be regarded as an upper bound of its potential impact, because our experiments have been performed in idealized settings, including the zonally homogeneous lower boundary condition and the drastic rise in subpolar SST with the removal of the SST front. A further modeling study with modest changes in SST distribution is needed for a deeper understanding of the role of oceanic frontal zones. It should be kept in mind that storm tracks and jet streams in the real atmosphere are influenced by many factors, including the land–sea thermal contrast and planetary waves forced orographically and thermally, in addition to SST gradients and STJs.

The results in this study illuminate the distinct characteristics of the two types of tropospheric westerly jets and associated storm tracks. A STJ is formed through angular momentum transport by the Hadley cell (Held and Hou 1980; Lindzen and Hou 1988). A STJ in our experiments accompanies no well-defined baroclinic zone at the surface (Fig. 2), as observed by NS04 in the SH winter. A storm track can be formed along a STJ, but
confined to the upper-level jet core with little baroclinic eddy growth in the lower troposphere. This storm track is maintained by the influx of eddy activity from a midlatitude baroclinic zone (Fig. 2), as observed (NS02; NS04), because an intense STJ core acts as a waveguide favoring the propagation of transient eddies. Significant baroclinic eddy growth would be possible along a STJ if the underlying SAT gradient was significant, but in reality the underlying SST gradient is rather weak (Fig. 1). In contrast, a PFJ is basically “eddy driven” and is maintained by the convergence of eddy momentum fluxes (Fig. 7; LK03; NS04). The PFJ and associated storm track display deep structures that are consistent with vigorous baroclinic eddy growth, despite modest baroclinicity in midlatitudes above the PBL relative to that below the wintertime STJ core. The Eady growth rate evaluated for such a lower-tropospheric layer as that between the 700- and 850-hPa levels has been used conventionally as a measure of mean-flow baroclinicity (e.g., Hoskins and Valdes 1990), but the results above suggest that near-surface baroclinicity is a more relevant measure of the mean-flow condition for baroclinic eddy growth, as argued by recent studies (e.g., Inatsu et al. 2003; NS04; Nakamura et al. 2004). In fact, near-surface baroclinicity has certain dynamical relevance because it corresponds to potential vorticity gradient at the surface, which is essential for baroclinic instability (Hoskins et al. 1985).

The cross-frontal differential sensible heat supply from the ocean is essential for retaining intense near-surface baroclinicity, as is observed along the midlatitude oceanic frontal zones (Nakamura et al. 2004, 2008; Taguchi et al. 2009; Nonaka et al. 2009; Hotta and Nakamura 2009, manuscript submitted to J. Climate). The restoring effect by an oceanic frontal zone on near-surface atmospheric baroclinicity is illustrated schematically in Fig. 12, as inferred from the lag relationships indicated in Fig. 10. The tight SAT gradient across a SST frontal zone leads to the development of baroclinic eddies (Fig. 12a). Vigorous poleward eddy heat transport relaxes the SAT gradient to yield greater SST–SAT differences (Figs. 10a,b), giving rise to enhanced and reduced
surface SHF\textsuperscript{6} at the equatorward and poleward flanks of the SST front, respectively (Figs. 10c,d). This enhancement of the differential heat supply from the ocean effectively restores the near-surface baroclinicity (Fig. 12b). In fact, the differential heat supply is greatly enhanced in the presence of an oceanic frontal zone in our experiments (Figs. 8a,b), as observed in the south Indian Ocean (Figs. 8g,h; Oberhuber 1988) and reproduced

\textsuperscript{6} On the cooler side of the SST front, the relaxed SAT gradient may lead to stronger PBL cooling by the ocean.
with a high-resolution coupled GCM (Nonaka et al. 2009) and a regional atmospheric model (Taguchi et al. 2009). This effective restoration of baroclinicity, referred to as “oceanic baroclinic adjustment” by Nakamura et al. (2008), differs fundamentally from the conventional view of the maintenance of baroclinicity in which differential radiative heating between high and low latitudes restores the meridional temperature difference (e.g., Stone 1972).

An oceanic frontal zone along the boundary between the subtropical and subpolar gyres (i.e., a subarctic frontal zone) or along the Antarctic Circumpolar Current (i.e., the APFZ) is fairly robust on seasonal and interannual time scales (Nakamura and Kazmin 2003; NS04; Nakamura et al. 2004), sustaining the differential heat release from the ocean. The robustness of the sharp SST gradient is due to the large thermal inertia of the deep ocean mixed layer and the maintenance by differential thermal advection between the warm and cold currents in the frontal zones (Qiu 2000, 2002; Kelly and Dong 2004). These are the features unique to the midlatitude oceanic frontal zones, in contrast to the central and eastern portions of midlatitude ocean basins where SST is quite variable to atmospheric forcing. In this regard, our paradigm of extratropical air–sea interaction illustrated in Fig. 12 differs from the paradigm by Kushnir et al. (2002, their Fig. 5), where the low-frequency anomalies of the meridional SST gradient is assumed to be a thermal response of the ocean mixed layer to atmospheric forcing.

Our results suggest that due consideration on the role of the ocean may be necessary for a full understanding of the extratropical atmospheric general circulation. Midlatitude storm tracks and westerly jets have generally been studied in a framework of purely atmospheric dynamics. From a viewpoint of poleward heat transport in the climate system, however, subtropical oceanic gyres transport a greater amount of heat to midlatitudes than the atmospheric circulation, especially over the NH (e.g., Trenberth and Caron 2001). The transported heat is released around midlatitude oceanic frontal zones into the atmosphere intensively from such warm western boundary currents as the Agulhas Current in the SH and the Gulf Stream and Kuroshio in the NH. As indicated by Nonaka et al. (2009), Taguchi et al. (2009), and our experiments, the heat supply from the ocean around the frontal zone maintains an atmospheric baroclinic zone and thereby acts to anchor a storm track along which baroclinic eddies grow to transport heat into higher latitudes. These processes are referred to as “the passing of a heat transport ‘baton’ from the ocean to the atmosphere in the midlatitudes” by Kelly and Dong (2004). Given this relationship, the observed collocation of a storm track and associated PFJ with an oceanic frontal zone seems to be not fortuitous, but rather a manifestation of the oceanic influence on the atmosphere, as postulated in Nakamura et al. (2004, 2008).

Our modeling results are also supportive of the possibility of a feedback loop among a midlatitude oceanic frontal zone, the storm track, and the surface westerlies, as was postulated by Nakamura et al. (2004, 2008). Our experiments suggest that an oceanic frontal zone acts to anchor a storm track, and its activity maintains the surface westerlies through its westerly momentum transport. The surface westerlies, in turn, drive the ocean circulation and act to deepen the oceanic mixed layer in concert with vigorous storm activity, which may contribute to the maintenance of the oceanic frontal zone to make the feedback loop up. Further observational and modeling studies are needed to assess whether this positive feedback loop in the extratropical atmosphere–ocean system is actually operative.

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REFERENCES


CORRIGENDUM

TAKEAKI SAMPE
International Pacific Research Center, SOEST, University of Hawaii at Manoa, Honolulu, Hawaii

HISASHI NAKAMURA
Department of Earth and Planetary Science, University of Tokyo, Tokyo, Japan

ATSUSHI GOTO
Climate Prediction Division, Japan Meteorological Agency, Tokyo, Japan

WATARU OHFUCHI
Earth Simulator Center, Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan

Due to a printing error, the title page of Sampe et al. (2010) contained a logo misidentifying the special collection to which it belongs. In fact, Sampe et al. (2010) is part of the U.S. CLIVAR Western Boundary Currents special collection, not the U.S. CLIVAR Drought special collection.

The staff of the Journal of Climate regrets any inconvenience this error may have caused.

REFERENCE