The location of the midlatitude storm track in aquaplanet AGCMs

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To be submitted to \textit{Journal of Atmospheric Science}

February 12, 2010

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Abstract

The sensitivity of the midlatitude storm track and eddy driven wind to the sea surface temperature (SST) boundary forcing is studied over a wide range of climate regimes using both simple and comprehensive general circulation models over aquaplanet lower boundary conditions. Under the circulation regime that the eddy driven jet is strongly coupled with the subtropical jet of the Hadley cell, similar to the conditions of the present climate, the eddy-driven jet shifts monotonically poleward with both the global mean and the equator-to-pole gradient of the SST. Whereas the eddy-driven jet can have a reverse relationship to the gradient if it is well separated from the subtropical jet and Hadley cell boundary in a double-jet circulation regime.

A simple scaling is put forward to interpret the simulated sensitivity of the storm track/eddy driven jet position within the single-jet dynamic regime in both models. The rationale for the scaling is based on the observed large compensation between the horizontal (momentum flux) and vertical (heat flux) components of the eddy Eliassen-Palm activity flux in the upper troposphere. Thus, the location of the maximum eddy momentum convergence in the upper troposphere, and hence the eddy driven wind, may be predicted from the eddy heat flux coming out of the lower troposphere. The latter can be parameterized by a measure of baroclinicity whose latitudinal variation shows a linear relationship to preferential perturbation in the static stability at the equatorward flank relative to the poleward flank of the storm track. To the extent that the preferential static stabilization is deterministically constrained by and hence can be predicted from the given SST conditions through the first principles for the midlatitude stratification, one may, given SST perturbations, predict which way the storm track and eddy driven wind should shift with respect to a chosen reference climate state. The resultant anomaly-wise scaling turns out to be valid for both idealized and comprehensive models. By corollary, it
can be argued that the poleward shift of storm track found in the global warming simulations by fully coupled climate models may be attributed, at least partially, to the robust increase in the subtropical and midlatitude static stability.
1. Introduction

Most of the midlatitude continents are under the influence of the storm tracks and the dynamics of storm tracks has been the central theme of climate dynamics inspiring numerous studies (e.g., Hartmann 1974; Blackmon 1976; Blackmon et al., 1977; Hoskins et al., 1989; Hoskins and Valdes, 1990; Chang 2001; Chang et al., 2002; Bengtsson, et al., 2006; ). The storm tracks have a conspicuous seasonal cycle in their locations, especially in the Northern Hemisphere: it shifts equatorward in step with the jet stream from fall to midwinter, and then migrates poleward afterwards (Hoskins et al., 1989; Nakamura, 1992). In addition, at interannual time scales, ENSO can impact on the latitudinal position of the storm track and the jet stream, with the warm phase leading to an equatorward contraction (Robinson, 2002; Seager et al., 2003; Orlanski, 2005; L’Heureux and Thompson 2006; Lu et al., 2008; Chen et al. 2008; Brayshaw et al., 2008) and the opposite movement for the cold phase. The issue of the latitudinal position of the storm track and jet stream shift is particularly topical given that the poleward shift of the mid-latitude storm tracks is deemed to be one of the most robust features under global warming (e.g., Kushner et al., 2001; Yin, 2005; Lorenz and DeWeaver, 2007; Lu et al., 2008; Chen et al., 2008). Nevertheless, the mechanistic research on the storm track has hitherto been largely focused on the intensity (e.g., Chang, 2001; Nakamura, 1992; Straus and Shukla, 1988), and much is yet to learn as to the mechanisms control the location of the storm track.

Hoskins and Valdes (1990) perceived the storm track as being self-maintaining in the sense that storm track eddies in general are most vigorous downstream of the region of peak baroclinicity, and the mixing of temperature by eddies is relatively benign where is baroclinicity is largest; further, the enhanced baroclinicity over the storm track entrance region is actively maintained by condensational heating, which in turn is caused by the cyclones themselves. This
depiction of storm track as the result of a *symbiotic* relationship between the transient eddy activity and the large-scale flow (Cai and Mak, 1990; Robinson, 1991) makes it difficult to predict the position of the storm track from the knowledge of the mean wind or temperature structure. Such an approach has been attempted by many to explain the shift of eddy field with a measure of baroclinicity such as the Eady Growth Rate (e.g., Inatsu et al., 2003; Brayshaw et al., 2008). However in many cases this method may not reveal causality, but only reaffirms the *symbiotic* relationship between the transients and the mean flow field: the storm track will always ensure that the baroclinicity is locally maximum nearby. Another difficulty with any possible closure theory for storm track is related to the second moment formulation (variance of zonal and/or meridional winds, or covariance between wind and temperature perturbations) for the storm track, which requires higher order moments to form a tractable problem.

What we pursue in this study is not necessarily the rigorousness of theoretical treatment for the storm track, but the associated structure of meteorological variables, such as precipitation, temperature. It is well documented (also as experienced by people living in the west coast of American and Eurasian continents) that the variance and covariance of eddy properties, vorticity, vertical motion, humidity, temperature and precipitation all bear a distinct relationship to the jet stream, in both the Pacific and Atlantic basins, in both the Northern and Southern hemispheres, and in the annual cycle and interannual variability. As will be demonstrated later with the idealized model, the latitudes of the maxima of eddy statistics do vary in tandem with the mean wind. This allows us to use the eddy-driven westerly wind as proxy for the storm track and any predictive power obtained for the westerly wind should have important bearings on the other properties of the midlatitude storm track.
Therefore, it is the position of the eddy-driven westerly wind that we strike for in this study. It is beyond the scope of the present study to construct a closure theory for the storm track, instead we only strive for the primary dynamic factor that can be used to infer the movement of the proxy of storm track—*eddy-driven surface westerly* (EDSW). An important attribute of the dynamic factor is that it should be quantifiable from the given boundary conditions, so that through it one might be able to link the variation in position of the storm track/EDSW to any given condition for the problem. This exercise, which we call scaling, will be carried out for a suite of aquaplanet model simulations under prescribed SST boundary conditions. The whole endeavor is motivated by the fact that the best-measured quantity in both modern and paleo-climate studies is the surface temperature, and one may potentially tap into this resource by forming a heuristic framework that links surface temperature to the midlatitude mean flow and eddy statistics. Any success along this line may lead to some predictive power for the changes of storm track in the past and future.

The models and experiments will be detailed in Section 2. Model simulations on the locations of the storm tracks and EDSW will be reported in Section 3. Section 4 lays out the rationale of the scaling, which is to be verified in Section 5. Finally we conclude with summary and discussion on the possible implication in climate change and climate variability.

**2. Model Description and Simulations**

2a. *Idealized GCM*

The idealized GCM is an ice-free, land-free model (so called aquaplanet model) consisting of various simplified physical parameterizations coupled to a spectral dynamical core that solves the primitive equations (Frierson et al., 2006; Frierson 2007a). For longwave radiative transfer, a
gray scheme is used with fluxes that are a function only of temperature, and not water vapor or cloudiness. Water vapor itself is a prognostic variable, but there is no liquid water or clouds. Condensation and moist convection occurs with a simplified version of the Betts-Miller convection scheme (Betts, 1986; Betts and Miller, 1986; Frierson 2007a) and a grid-scale condensation scheme. The simplified model physics also includes a simplified Monin-Obukhov surface flux scheme, and a K-profile boundary layer scheme. All simulations in this study are run at a resolution of T42 in horizontal, and 25 levels in vertical.

2b. Full GCM

The full GCM simulations in this paper are the same simulations originally used to study poleward heat transports in the study of Caballero and Langen (2005). The model is a comprehensive GCM, with full representation of clouds, radiation, convection, and other physics. The atmospheric model used for these simulations is the Parallel Community Climate Model Version 3 (PCCM3), which is the atmospheric component of the Fast Ocean-Atmosphere Model (FOAM) (Jacob 1997). The model uses the physical parameterizations of the NCAR CCM3.6 model (Kiehl et al. 1996) and the dynamical core of the NCAR CCM2 model. The full GCM is run at T42 resolution, with 18 vertical levels. Sea ice is specified where SST is below 0°C.

2c. Boundary conditions

The boundary conditions used here are from the study of Caballero and Langen (2005). The surface is an aquaplanet (ocean-covered Earth) with no topography, and prescribed, zonally symmetric SST distributions. The SSTs are functions of latitude (\( \phi \)) only

\[
T_s(\phi) = T_m - \Delta T (3\sin^2 \phi - 1) / 3,
\]

(1)
controlled by two parameters: $T_m$ is the global mean temperature, $\Delta T$ is the equator-to-pole temperature difference. The functional form is chosen so that changes in $\Delta T$ result in no change in global mean temperature and the maximum gradient is always located at 45° latitude. We examine simulations with $T_m$ between 0 and 35°C, and $\Delta T$ between 10° and 60°C. For the idealized model, simulations are run at 10K increments for $T_m$ and $\Delta T$, with additional runs of $T_m=35°C$; for the full GCM experiments, simulations are conducted for 5K intervals except the few cases that the tropical SST is too warm to be relevant. It should be noted that due to the functional form of (1), the tropical temperatures increase with increases in both mean temperature and temperature gradient. All simulations are spun up for 1 year, and statistics are calculated over 3 subsequent years of integration. The time mean fields are calculated by averaging the Northern and Southern Hemispheres since the prescribed SST and the resulting model climatology are hemispherically symmetric.

3. Results of Model Simulations

We first examine the EDSW position in the suite of simulations with the two models. The EDSW, as the term implies, is a result of the balance between the vertically integrated eddy momentum convergence and the surface drag. Thus, intuitively, the maximum EDSW may be used as a proxy indicator for the storm track axis. Indeed, the axis of storm track (measured as vertically integrated Eddy Kinetic Energy, EKE) is in excellent alignment with the latitude of the maximum EDSW in the idealized model simulations (Figure 1c). To a lesser extent, this relationship also holds for the full GCM (Figure 1d). In general, it is justifiable that the shift of the EDSW is representative of that of the EKE of the storm track, with some cautionary
discretion applied to the full model. This result, in a context of our aquaplanet simulations, corroborates the notion that mean wind moves in concert with the transients.

Both the idealized and full GCMs simulate a monotonic poleward shift of the EDSW/storm track with increasing global mean temperature $T_m$—probably the most robust feature in all the simulations. Another robust feature, not shown, is the intensification of storm activity and the EDSW with increasing temperature gradient $\Delta T$ (see Figure 3a in Caballero and Langen, 2005). However, the two GCMs show distinct difference in the sensitivity of the storm track/EDSW position to $\Delta T$. For the idealized model, the position of EDSW is also a monotonic function of $\Delta T$, moving poleward with larger $\Delta T$. This is not the case for the full GCM. In the middle of the $T_m-\Delta T$ space, as $\Delta T$ increases from its lowest value, the EDSW first shifts equatorward, and then poleward after passing its most equatorial position at the intermediate value similar to the conditions of present climate. The distinct behavior in the two models precludes a common scaling theory for the position of storm track/EDSW that is valid for both models. This is why this study will only be focused on an anomaly-wise scaling for the anomalous shift of EDSW with respect to a certain chosen reference state.

Further inspection on the zonal wind profiles of the full AGCM implies that this peculiar sensitivity of equatorward shift with $\Delta T$ may be related to the large separation between the subtropical jet and eddy-driven jet when the mean temperature is warm but the gradient is weak (the upper left domain in Figure 1b). The cases where wind is characterized by a distinct double-jet condition are marked with crosses in Figure 1b. It is worth noting that the wind profiles in the double-jet regime bear great resemblance to those of the bifurcation discussed by Lee and Kim (2003) and Chen and Plumb (2009). In the case of Lee and Kim (2003), the bifurcations occurs when the subtropical jet is intensified by a equatorial heating and as a result, the midlatitude
disturbances are more subject to the organizing effect of a stronger subtropical jet (which itself is stable in position) and hence shift equatorward. The sharp regime transition toward a merged jet at ~30° latitude never takes place in our case, instead, the eddy-driven jet merely progresses gradually equatorward as the subtropical jet grows in strength with increasing $\Delta T$. As one approaches the middle value of $\Delta T$, the eddy-driven jet and the subtropical jet start to merge, and a different mechanism begins to come into play in the response of EDSW position to $\Delta T$ — a subject to be elaborated in the following sections. A typical transition from double- to single-jet regime in the full GCM with increasing SST gradient $\Delta T$ is exemplified by Figure 2, which shows the profiles of 400 hPa zonal wind for cases of $T_m=15^\circ C$, with $\Delta T$ varying from 10 to 50K at increments of 5K.

4. Rationale for scaling

As mentioned before, the scaling framework in this section is built for the location of the mid-latitude surface wind. Given the *symbiotic* relationship between the EDSW and the eddy activity, any success with the scaling for the position of EDSW should have immediate implication for the storm track as well.

The scaling framework is built upon the understanding derived from the zonal momentum balance in the two-layer QG model: the meridional structure of the lower tropospheric wind is shaped by the vertically integrated eddy potential vorticity flux (Robinson 2000, 2006; Pavan and Held, 1996), i.e.,

$$\bar{u}_2 = \frac{1}{\kappa} (\mathcal{F}_1 + \mathcal{F}_2)$$

(2)
where the overbar denotes zonal mean; \( \kappa \) represents the rate of a frictional damping on the lower level velocity; \( \mathcal{F}_i = (v'_i q'_i) \) are the zonally averaged eddy fluxes of QG potential vorticity with subscript \( i=1, 2 \) indicating the upper and lower troposphere, respectively. Under the QG approximation, these are equivalent to the divergence of the Eliassen-Palm (EP) fluxes on both levels (Edmon et al., 1980), so that \( \mathcal{F}_2 \), which is typically positive in the real atmosphere, may be considered the source of Eliassen-Palm eddy activity, while \( \mathcal{F}_1 \) may be considered its dissipative sink aloft. To apply this framework to a continuously stratified atmosphere, we use the vertical integration over levels above and below 560 hPa to estimate the eddy fluxes for the upper (1) and lower (2) troposphere, respectively. With the atmosphere partitioned as such, and the QG Eliassen-Palm flux defined as

\[
\mathbf{F} \equiv (-u'v') \mathbf{j} + \left( \frac{f_0 v' \theta'}{\theta_p} \right) \mathbf{k},
\]

the eddy PV fluxes (\( \mathcal{F}_i \)) and its components of associated momentum flux convergence (\( \mathcal{M}_i \)) and heat fluxes (\( \mathcal{H}_i \)) can be expressed as

\[
\mathcal{F}_1 \equiv \frac{1}{p_s} \int_{p=0}^{p=560} \mathbf{\nabla} \cdot \mathbf{F} \, dp = \mathcal{M}_1 + \mathcal{H}_1 \tag{3a}
\]

\[
\mathcal{F}_2 \equiv \frac{1}{p_s} \int_{p=p_s}^{p=560} \mathbf{\nabla} \cdot \mathbf{F} \, dp = \mathcal{M}_2 + \mathcal{H}_2 \tag{3b}
\]

where
\[ \mathcal{M}_1 = \frac{1}{p_s} \int_{p=0}^{p=560} dp \left\{ \frac{\partial}{\partial y} (-u'v') \right\}, \]

\[ \mathcal{H}_1 = \frac{1}{p_s} \int_{p=0}^{p=560} dp \left( \frac{f_0 v'v'}{\theta p} \right) \right|_{p=0}; \]

\[ \mathcal{M}_2 = \frac{1}{p_s} \int_{p=p_s}^{p=560} dp \left\{ \frac{\partial}{\partial y} (-u'v') \right\}, \]

\[ \mathcal{H}_2 = \frac{1}{p_s} \int_{p=p_s}^{p=560} dp \left( \frac{f_0 v'\theta'}{\theta_p} \right) \right|_{p=0}. \]

Boundary condition of no perturbation has been applied to \( p = 0 \) in the expression of \( \mathcal{H}_1 \) and to \( p = p_s \) in the expression of \( \mathcal{H}_2 \). Observations (Edmon et al., 1980) demonstrate that the meridional structure of the vertically integrated PV flux (\( \mathcal{F}_1 + \mathcal{F}_2 \)) near the latitude of the maximum surface wind is dominated by its low level fluxes (\( \mathcal{F}_2 \)), and which is in turn dominated by the heat flux component (\( \mathcal{H}_2 \)) of the Eliassen-Palm fluxes. Figure 3 depicts the meridional profiles for upper tropospheric PV flux \( \mathcal{F}_1 \) (blue), upper tropospheric momentum flux convergence \( \mathcal{M}_1 \) (black dashed), lower tropospheric PV flux \( \mathcal{F}_2 \) (red), and lower tropospheric eddy heat flux \( \mathcal{H}_2 \) (red dashed), all normalized by the maximum of \( \mathcal{F}_2 \), and the surface wind (black) for the full suite of simulations with the idealized model. With increasing SST gradient and hence increasing intensity of the transient activity, \( \mathcal{F}_2 \) becomes more dominant over \( \mathcal{F}_1 \) and the meridional structure of the surface wind tends to be increasingly shaped by \( \mathcal{F}_2 \), the peak of which is dictated by that of \( \mathcal{H}_2 \) (as one can see by comparing the two red curves in Figure 3), while the momentum flux in the lower troposphere is negligible. It is worth noting that across all the SST cases there seems to be a tendency for the peak of the upper level momentum flux convergence \( \mathcal{M}_1 \) (hence the surface wind) and the peak of \( \mathcal{H}_2 \) to move in the same direction in
their variations, although their peaks do not entirely overlap. By corollary, this implies a
tendency for the EP eddy activity flux (predominantly $\mathcal{H}_2$) entering from below into the upper
troposphere to radiate away from the source latitude thereby maintaining the total PV flux to be
small near the source. For the real atmosphere, this is manifested as the weak positive eddy PV
flux near the core of the eddy driven jet (as delineated, for example, in Figure 12.19 in Vallis
(2006)). In principle, the PV flux should vanish under the nonacceleration conditions of a steady
inviscid adiabatic flow (Andrews and McIntyre 1978a,b; Edmon et al. 1980). The possible
reason for the minimum total PV flux near the EDSW is that the large environmental PV
gradient in the upper troposphere provides a condition favorable for wave-like propagation than
irreversible PV mixing. This is essentially the same situation as discussed in DelSole (2001)
wherein the eddy momentum fluxes are reproduced as a response to the lower tropospheric eddy
PV stirring. For an extreme case where the eddy PV forcing is confined entirely at the lower
boundary, a scenario discussed by Robinson (2000), the barotropic wind should have exactly the
same meridional structure as the surface heat/PV flux (the component associated with vorticity
flux, $\mathcal{M}_v$, is neglected). For this case, the heat flux ($\mathcal{H}_h$) component is completely compensated
by the vorticity flux ($\mathcal{M}_v$) component in the upper level, resulting in zero net PV forcing there.
To the extent that this compensation is operating in the response to the prescribed SSTs in our
simulations, the latitude of the EDSW could be considered to be set by the latitude of the
maximum low-level heat flux, i.e.,

$$\phi_{\partial H_{\nu}/\partial y=0} \approx \phi_{\partial H_{r}/\partial y=0},$$  

(4)

alternatively,

$$\partial \ln \bar{u}_2 / \partial y \sim \partial \ln H_2 / \partial y \quad \text{near} \quad \phi_{\partial H_{\nu}/\partial y=0},$$  

($4'$)
Obviously, the agreement between the locations of the EDSW and maximum \( \mathcal{H}_2 \) is not excellent for all SST cases since the compensation between eddy vorticity flux and heat flux in the upper layer is far from perfect. Notwithstanding, the fact that the compensation occurs at all should be capitalized on to predict the shift of EDSW/storm track, and it does turn out to bear fruit when we try to exploit the compensation mechanism in an anomaly-wise way (elaborated as follows). How we construct this is described below.

First of all, since the maximum of \( \mathcal{H}_2 \) is indicative of the position of the EDSW in either the mean states or the deviation from a reference state, we attempt to parameterize \( \mathcal{H}_2 \) with the zonal mean thermal structure so that we may express the zonal index variability of the wind in terms of the zonal mean temperature. Given that the lower tropospheric finite amplitude eddy production is fundamentally local and hence effectively diffusive, the eddy-induced heat flux may be related to the local mean temperature gradient in form of \( \mathbf{v}'\theta' = -D \frac{\partial \bar{\theta}}{\partial y} \). With no much discretion given on the exact form of its power law relation to \( \partial \bar{\theta} / \partial y \) and \( \partial \bar{\theta} / \partial p \), we simply assume the diffusivity coefficient to be a constant and use the vertically integrated mean gradient in (5). As a result,

\[
\left( \mathbf{v}'\theta' \right)_{p_{so}} \sim -D \left( \frac{\partial \bar{\theta}}{\partial y} \right),
\]

where the angle bracket denotes vertical integration over the troposphere (from surface to tropopause). This simple relation turns out to be a better fit to the model results than more nuanced choice of diffusivity parameterizations (such as those proposed by Green (1970), Stone (1972), Held (1978a), and Branscome (1983)). Nevertheless, the final result regarding the position of the storm track/EDSW is not very sensitive to the specific choice of parameterization
of eddy heat flux. Further approximating the stratification at 560hPa with the tropospherically
averaged one yields

\[ \mathcal{H}_2 \propto -f_0 \left( \frac{\partial \bar{\theta}}{\partial y} \right) \left( -\frac{\partial \bar{\theta}}{\partial p} \right)^{-1} \equiv \xi, \tag{6} \]

The modulating effect of the factor \( 1/p_s \) on the meridional structure of \( \mathcal{H}_2 \) has also been
neglected in (6). The right hand of (6) measures approximately the isentropic slopes in a vertical
average sense, which is also an indicator for baroclinicity. We denote the right hand side of (6)
as \( \xi \), thus if this simple parameterization is valid one should expect the axis of the storm
track/EDSW varies in concert with the maximum of \( \xi \). This is indeed the case as shown in
Figure 4a and 4b for both the storm track (EKE) and EDSW for the idealized model. The full
model simulations show a similar agreement (not shown for brevity). There is an approximate 4°
offset between the storm track/EDSW axis and the maximum baroclinicity, a subject we will
return to in next section. It is important to note that a similar fit results when the actual peak
latitudes of \( \mathcal{H}_2 \) are plotted, indicating that \( \xi \) can successfully represent the profile of \( \mathcal{H}_2 \) in a
parameterization-al sense. With this link established, if one were able to predict the shift of \( \xi \),
this should lead to the information about the shift of storm track/EDSW.

Since it is the shift of the EDSW with respect to the well observed reference climate (for
example, the modern climate), not the absolute location, that is the point of interest of this study,
next we choose the case of \([T_m=20^\circ C, \Delta T=40^\circ C]\) (the central case in Figure 2), a condition
typical of the present day climate, as the reference state and treat all the information about this
state as known and all other cases as deviation from it. Denoting \( y_0 \) as the location of the
maximum EDSW of the reference state, the discrete form of equation (4’) about \( y_0 \) with \( \mathcal{H}_2 \)
being replaced by \( \xi \) becomes
Here, $U$ represents $\bar{u}_2$ simply for notational reason; the superscript $+$ (-) indicates a latitudinal average over the polar (equatorial) flank of the reference latitude $y_o$ (see Figure A1 for illustration); subscript $m$ indicates the maximum value near $y_o$. In practice, we choose two latitude bands ([27°-47°] and [47°-67°]) centered around $y_o = 47°$. This is equivalent to discretizing (4') on a very coarse 20° grid. Differential perturbations of the wind speed on either sides of $y_o$ should lead to a shift, which, with the aid of a Taylor series expansion, can be expressed as

$$\delta y \approx \frac{U_r m}{U_p m} \cdot \left( \delta U^+ - \delta U^- \right) \cdot \left\{ \left[ \frac{dU_r}{dy} \right]^+ - \left[ \frac{dU_r}{dy} \right]^+ \right\}^{-1},$$

where, $U_r$ and $U_p$ are the reference and perturbed surface wind, respectively; $\delta U^{+(-)}$ indicates the wind perturbation averaged within the [47°-67°] ([27°-47°]) latitudinal band; the subscript $m$ indicates the amplitude at the wind axis. As a result, the shift of the jet under perturbations in both the latitude and the magnitude is not only proportional to the differential wind change, but also inversely proportional to the amplitude of the perturbed wind. See Appendix A for the detail of derivation of (8). After a trivial manipulation, (8) can be approximated as

$$\delta y \approx \frac{U_{rm}}{dU_r}{dy} - \frac{dU_r}{dy} \cdot \left[ \frac{U_p^+ - U_p^-}{U_{pm}} - \frac{U_r^+ - U_r^-}{U_{rm}} \right]$$

$$\left[ \frac{dU_r}{dy} \right]_+ - \frac{dU_r}{dy} \cdot \left[ \frac{U_p^+ - U_p^-}{U_{pm}} - \frac{U_r^+ - U_r^-}{U_{rm}} \right],$$

$$\left(9\right)$$
Insofar as the perturbation is not too large to invalidate the relation (7), one may substitute (7) into (9) and obtain a proportional relationship between the shift of the EDSW and the fractional change of $\xi$

$$\delta y \propto \frac{\delta \xi^+ - \delta \xi^-}{\bar{\xi}_{pm}}, \quad (10)$$

The way this relation is derived renders itself to be valid only for small perturbation about certain reference state. However, to test the limit of it, we examine the right hand side of (10) for all the cases. Furthermore, instead of relating the latitude of EDSW to the latitude of maximum baroclinicity, our scaling links the latitude of EDW to the differential change in baroclinicity.

5. Verification

5a. Relationship between the axis of storm track and maximum $\xi$

The latitudinal locations of maximum $\xi$ are plotted against the locations of the storm tracks and the maximum surface wind in Figure 4a and 4b, respectively. Despite a tendency for the maximum $\xi$ to be located on the poleward side of the axis of the storm track and the surface wind, the linear relationship between each other’s variation is clear. While this should not be interpreted as evidence for a cause-effect relationship, it does support the notion of the “symbiotic” existence of the transient eddy activity and the mean flow. The $\sim 4^\circ$ offset between the location of the storm track/EDSW and the maximum of $\xi$ should not negate the rationale based on relation (4), since it can be easily reconciled by introducing an offset into this relation with the subsequent reasoning about the perturbations remaining valid.

Given the linear relationship between shift of $\xi$ and the storm track/EDSW, we next examine how $\xi$ and its components, i.e., the static stability and the tropospheric temperature
(meridional) gradient, vary under each SST condition. Figure 5a shows the breakdown of the profiles of $\xi$ for all the cases. Overall, the magnitude of $\xi$ increases with $\Delta T$, and decreases substantially with $T_m$ owing to the increase of static stability. A similar tendency is also found in the simulations with the full GCM. As a result, the mid-latitude isentropic slope flattens considerably with increasing mean SST, in contrast to the constant isentropic slopes as rationalized from the arguments of baroclinic adjustment (Stone, 1978; Stone and Nemet, 1996) or neutral supercriticality owing to weak nonlinearity in the atmospheric eddy-eddy interactions (Schneider and Walker, 2006). This result suggests that the existing dry theories for the midlatitude adjustment might be inadequate to describe the mid-latitude tropospheric thermal stratification under large SST boundary perturbations, as least for these idealized experiments examined here (Juckes, 2000; Frierson, 2008; Schneider and O’Gorman, 2008). In the meantime, the flattening of the isentropic slope with increasing mean temperature is consistent with previous studies on the effects of moisture on baroclinic eddies (Stone and Yao, 1990; Held, 1978b; Gutowski et al., 1992). Held (1978b) studied the effect of adding a hydrological cycle to a dry two-layer climate model, and calculated the ratio of the mean vertical eddy heat flux, weighted by the static stability, to the meridional eddy heat flux weighted by the meridional gradient of potential temperature—a ratio that approximates the mixing slope divided by the isentropic slope. Held found that this ratio was 0.55 in the dry model, close to the value given by classical theories of baroclinic instability, but increased to 0.90 in his moist case. The increase of the ratio of the mixing slope with respect to the mean isentropic slope was also found in Gutowski et al. (1992) to be the key effect of condensation in the life cycle of mid-latitude eddies. Given the fact that vertical transport of dry static energy by eddies ($w'\overline{\theta}'$) counters the background gradient ($\partial \overline{\theta} / \partial z$), it is conceivable that the increase of mixing slope may contribute
to the flattening of the isentropic slope as the uniform SST warming increases moisture content. At any rate, the involvement of moisture makes the maintenance of the midlatitude mean thermal structure an issue of three-way interplay among (i) the horizontal eddy heat flux; (ii) the vertical eddy heat flux; and (iii) the eddy-related diabatic heating, a subject remaining of great challenge for theoretical understanding.

The most prominent aspect of the static stability is its increase with $T_m$ as shown in Figure 5b for the idealized model (see also Frierson, 2008). A qualitatively similar result is also found for the full model, thus not shown for brevity. The increase is most prevalent in the tropics, where the vertical thermal profile approximately follows the moist adiabat. The increase also spreads poleward outside of the territory controlled by Hadley cell. From the coldest and flattest SST case to the warmest and steepest SST case, the static stability averaged over near the edge of the Hadley cell or the equatorward flank of the reference storm track (approximately between 25° and 45°) increases by a factor of 8. For each group with fixed $\Delta T$, the static stability increases preferentially over the subtropical latitudes relative to higher latitudes, pushing the peak of $\xi$ poleward, a phenomenon that is typical of the response to the GHG induced global warming (Frierson, 2006; Lu et al., 2008; Yin, 2005).

The tropospheric temperature gradients in the simulations are depicted in Figure 5c, together with the SST gradient (black curves). For the warm $T_m$ cases, the gradient profiles exhibit two peaks, a subtropical one associated with the thermally-forced subtropical jet (a same suite of simulations running on an axisymmetric configuration elucidate the subtropical peak unambiguously, not shown), and a subpolar one associated with the eddy-driven jet. For the intermediate and weak $T_m$ cases, these two peaks are whereas merged and indistinguishable. If one defines the edge of the Hadley cell as the subtropical peak of the temperature gradient, the
width of the Hadley cell can hardly exceed 35° latitude, in accordance with the notion that the scale of the Hadley cell is set by the terminus of the angular momentum conservation regime by the baroclinic instability (Held, 2000; Lu et al., 2007, Frierson et al., 2007b). However, the location of the EDSW associated with the eddy momentum convergence varies over a much wider range between 30° and 65° latitude (see also Figure 1). It is particularly notable that, for the cases with large $T_m$ and $\Delta T$, the EDSW maximum and the associated maximum temperature gradient appear significantly displaced on poleward side of the maximum SST gradient (located at 45° for all simulations). This displacement reflects the notion on the self-maintaining eddy-driven jet: a self-maintaining jet preserves and sometimes reinforces the mid-latitude gradient and place the gradient at poleward side of the imposed baroclinicity through the Transformed Eulerian Mean (TEM) overturning circulation (Robinson, 2000; Robinson, 2006; Chen and Plumb, 2009). The creation of baroclinicity in the life cycle of eddies was demonstrated clearly in the seminal work of Simons and Hoskins (1978) and Hoskins (1983). The self-maintaining nature of the EDSW prevents us from predicting the storm track or jet position based on the information of the atmospheric temperature gradient, since which itself is partially the result of the baroclinic eddy adjustment. Notwithstanding, one can still acquire some key information regarding the storm track from the static stability, through not only its direct contribution to the parameter $\xi$, but also its impingement on the eddy feedback to the mean temperature gradient. One can imagine an extreme paradigm in which the eddy adjustment that reshapes the mean thermal structure be totally submissive to the control of the subtropical static stability, the variations of the location of the storm track should fall into a straight line with the changes in the static stability. In the following, we will assess to what extent this paradigm is operating in the shift of the storm track/EDSW in these slab aquaplanet simulations.
5b. Variation about a reference state

Here, we choose the case \((T_m = 20, \Delta T = 40)\) as a reference state that is typical of the parameters for the present-day climate. Note that the following result is not very sensitive to which exact pixel point around \((T_m = 20, \Delta T = 40)\) in the \(T_m - \Delta T\) space is chosen as the reference state. We then compute \(\frac{\delta \xi^+ - \delta \xi^-}{\xi_{pm}}\) with the denominator estimated at the latitude of the reference wind maximum and the numerator at its flanks. The result is shown in Figure 6, reflecting a generally linear relationship between the shift of the EDSW and the fractional change of \(\xi\) for a certain range of SST perturbations, and a deterioration of the linearity when \((T_m, \Delta T)\) values deviate markedly from the reference value. This corroborates the anticipated link between the differential change in the baroclinicity (and hence the upward eddy EP fluxes) and the meridional distribution of upper tropospheric generation and dissipation of eddy activity for small perturbations. For example, if the baroclinicity increases on the poleward side of the storm track relative to the equatorward side, the source of the baroclinic wave activity shifts poleward. These anomalous waves are probably able to propagate away from their source latitudes owing to the waveguide effect of the jet, the momentum flux in the upper troposphere will change accordingly giving rise to a poleward shift of EDSW.

5c. Implication from a static stability theory

It now seems viable to predict the change of \(\xi\) based on the basic principles that link it to the given SST conditions. Through this exercise, we may obtain some predictive power about the shift of the storm track/EDSW given the SST boundary condition. Given the nonlinear nature of the midlatitude dynamics, it is elusive (at least to the best of our knowledge) to predict \(\xi\) per se from the SST. However, the bulk static stability over the storm track region appears to be
strongly constrained by the distribution of the boundary layer moist entropy and there has been some success in predicting the former based on the latter (Frierson, 2006, 2008) through the mechanism proposed by Juckes (2000). In the following part of the section, we simply predict the static stability component of $\zeta$ in $\frac{\delta \zeta^+ - \delta \zeta^-}{\zeta_{pm}}$ following Frierson (2008) to see how much leverage we can get on the shift of storm track from the knowledge of static stability alone.

The theoretical estimation for the bulk (dry) static stability is derived based on the simple scaling of Juckes (2000) and the variant thereof (Frierson 2006; 2008) for the mid-latitude moist stability. This theory relates the bulk moist stability $\Delta \bar{\theta} = \theta^* - \theta^s$, defined as the difference of the tropopause saturation equivalent potential temperature minus the equivalent potential temperature near surface (with * indicating saturation, and subscripts $t$ and $s$ tropopause and surface, respectively), to the meridional gradient of surface equivalent potential temperature through a mixing length closure:

$$\theta^* - \theta^s \sim L \partial_y \theta^s$$

(11)

where $L$ can be interpreted as the typical width of the storms. The recapitulation of this theory is provided in detail in Appendix B. In practice, we use the equivalent potential temperature difference between two fixed latitudes (between 30° and 50°, this is equivalent to fixing $L$ in (11)) to estimate the bulk moist stability, i.e., $\Delta \bar{\theta} \sim \theta^t_{30^\circ} - \theta^s_{50^\circ}$ at the equatorward flank of the reference jet position. To derive the dry stability $\Delta \bar{\theta} = \bar{\theta} - \bar{\theta}^s$ for the 30°-50° band, first, we compute from the SST the potential temperature ($\bar{\theta}$) and equivalent potential temperature ($\bar{\theta}^s$) near 40° latitude, assuming a constant relative humidity 0.8; then the near-tropopause level (300 hPa is actually used) saturation equivalent potential temperature should be $\bar{\theta}^*_t \equiv \bar{\theta}^s + \Delta \bar{\theta}_t$; last,
the upper level potential temperature $\bar{\theta}$ can be retrieved through the relationship between potential temperature and saturated equivalent potential temperature at the given pressure (300 hPa). The resultant estimate of the subtropical bulk static stability corresponding to 40° latitude is shown (Figure 7) to be in excellent agreement with the actual simulations.

In view of that the static stability does not vary as much over the poleward side as compared with the equatorward side of the jet (Figure 5b), we neglect the contribution of $\delta \xi^+$, and approximate $\delta \xi^-$ with $\delta \xi_l^-$, in the evaluation of $\frac{\delta \xi^+ - \delta \xi^-}{\xi_{pm}}$. As such, $\frac{\delta \xi^+ - \delta \xi^-}{\xi_{pm}}$ is simplified to be $-\frac{\delta \xi_l^-}{\xi_{pm}}$, where $\delta \xi_l^-$ is computed using the reference tropospheric temperature gradient and theoretical estimate of static stability at the equatorial flank of the jet. The result is presented in Figure 8, showing a tight relationship between this estimated quantity and the axis of the storm track with a correlation coefficient at 0.98. This strongly suggests that eddy activity and the eddy-induced wind do feel the changes in subtropical static stability, with an increase in latter causing a poleward shift of the former. In particular, a unit increase in $-\frac{\delta \xi_l^-}{\xi_{pm}}$ can lead to ~7° shift of the eddy-driven jet (or storm track). Any scaling that really works should not dependent on the model configurations. Next, to test the issue of model dependency, we apply this same scaling to the simulations with the full AGCM over the same suite of SST boundary conditions.

5d. Application to full AGCM

Again, as for the idealized model, we choose the case ($T_m = 20$, $\Delta T = 40$) as the reference. The result of applying the same scaling to all cases with the full model turns out to show no simple
linear relationship between $-\frac{\delta \xi_t}{\xi_{pm}}$ and the actual shift (Figure 9a). This may be related to the nonlinear regime behavior of the position of the EDSW over the wide range of $\Delta T$ as discussed in Section 3. Indeed, when applying to the $T_m-\Delta T$ domain wherein the wind profiles are mostly characteristic of single-jet or intermediate jet conditions and the EDSW shows a monotonic relationship to both $T_m$ and $\Delta T$, the linear relationship between $-\frac{\delta \xi_t}{\xi_{pm}}$ and the actual shift still holds markedly well (Figure 9b). It is even more encouraging to note that the slope of the linear fit is not that different from the idealized model — ~6° shift for per unit increase of $-\frac{\delta \xi_t}{\xi_{pm}}$.

Meanwhile, the linearity appears to be robust under different parameterizations of the eddy heat flux, but the slope of the linearity does not (not shown). This implies that, given the structure of the SST, we may be able to predict which way the storm track/EDSW should shift, but not how much, owing partly to the lack of theoretical understanding of eddy fluxes in a moist atmosphere. On the positive side, the substantial importance of the static stability in moderating the location of the storm track is clear.

6. Summary and concluding remarks

The mid-latitude storm track shifts poleward in the IPCC climate model simulations under the GHG-induced global warming. No widely accepted theory exists to rationalize this phenomenon, except some studies suggest that it links to the rise of the tropopause (Williams, 2006; Lorenz and DeWeaver, 2007) and the attendant eddy feedbacks through the eddy phase speed changes (Chen and Held, 2007; Chen et al., 2008; Lu et al., 2008). Here, by diagnosing a
suite of idealized aquaplanet simulations with specified SST boundary conditions, we demonstrate that the variation of the zonal index (or the shift of the surface wind) is constrained by the structure of the lower tropospheric Eliassen-Palm activity flux (primarily the heat flux) coming out of the lower troposphere. Further, the latter, parameterized as a quantity measuring the tropospheric baroclinicity, shows a strong sensitivity to the perturbation in subtropical static stability. In both an idealized GCM and a GCM with full physics, the preferential stabilization on the equatorial flank of the storm track/EDSW can, at least within the climate regime similar to the present climate, shift poleward the baroclinicity and hence the production of the eddy activity flux. Since the divergence of the horizontal eddy activity flux (equivalent to the convergence of eddy momentum flux) dictates the position of the EDSW through the momentum balance upon vertical integration, a poleward displacement of eddy heat flux portends a poleward shift of EDSW under the non-acceleration assumption for the zonal flow that anomalous wave generated at the surface can propagate away from the latitude of wave generation. Through this chain of reasoning, the displacement in the position of EDSW may be predictable provided the information of static stability can be derived exclusively from the given condition (SST for this case), particularly in view of that theory for the static stability work remarkably well for its prediction given SST in both the idealized and the full GCMs (Frierson, 2008). Finally, a good relationship results in both the idealized and full GCMs between the theory-predicted subtropical static stability and the position of EDSW, suggesting the static stability as an important leverage for the shift of storm track and the associated eddy-mean flow interaction.

Such a suggestion is additionally corroborated by the analysis of the simulations of the state-of-the-art CMIP3 models: as one can infer from Figure 6b in Lu et al. (2008), the models with greater increase in static stability to the immediate equatorward side of the climatological mean
EDSW tends to show a larger poleward displacement in EDSW. This rendition should not be considered contradictory, but complementary to the other mechanisms that emphasize the changes near the tropopause (e.g., Lorenze and DeWeaver, 2007). For the case of global warming, the upper tropospheric warming and stratospheric cooling can affect upper tropospheric eddy momentum flux through changes in the upper tropospheric wind and eddy phase speed (Chen et al. 2007; Chen and Held, 2007; Chen et al., 2008). Although the relative contributions to the EDSW shift from the boundary forcing and the upper troposphere-lower stratosphere remain to be quantified, it is clear that what determines eddy momentum flux, either through the non-acceleration assumption of zonal wind as discussed here, or eddy phase speed, or wave refraction, is the key to explain the wind shift.

Since this scaling is valid for perturbations in both the global mean and the latitudinal gradient of the SST, it may have some bearings on the change of the storm track in the past of the Earth’s climate. For example, during the glacial periods when the global mean temperature was about 10 degrees cooler than the present, the theory would suggest the storm track would be several degrees equatorward relative to today. This speculation is consistent with the higher-than-today loading of mineral dust from ice core records (e.g., Petit et al., 1999), attributable to the larger exposure of the subtropical desert to the uplift by midlatitude storm systems as they retreat equatorward in a colder climate (Chylek et al., 2001). During the mid-Holocene (about 6000 yrs ago), the solar radiation in southern hemisphere had a weaker equator-to-pole gradient during the austral spring followed by an overall dimming during summer, which could conceivably lead to an equatorward movement of storm track in southern summer. Whether this is true remains to be verified by paleoclimatic data.
Appendix A: Formulation of shift

Empirically speaking, dipolar wind anomalies centered at the axis of the jet can lead to a shift of the jet. Here, we derive a functional relationship between the shift of the jet and the dipolar wind anomalies considering perturbations in both the latitudinal position and the magnitude of the jet but with the shape kept the same.

Figure A1 Schematics for the shift of the jet under perturbations in both the axis and the amplitude of the jet.

We denote the reference and perturbed wind profiles as $U_r$ and $U_p$, respectively, and the corresponding maximum of the jet as $U_{rm}$ and $U_{pm}$. Contrasting $U_p$ to $U_r$, the difference, with the aid of a Taylor expansion, can be expressed as
\[
\delta U \equiv U_p - U_r \\
= (1 + \alpha)U_r (y - \delta y) - U_r (y) \\
= (1 + \alpha) \left\{ U_r (y) - \frac{dU_r}{dy} \cdot \delta y \right\} - U_r (y) \\
= \alpha U_r (y) - \left( 1 + \alpha \right) \left\{ - \frac{dU_r}{dy} \cdot \delta y \right\} 
\]

where \( 1 + \alpha = \frac{U_{pm}}{U_{rm}} \), and \( \alpha \) is the amplification factor with respect to the reference magnitude.

Averaging \( \delta U \) within the two latitude boxes at the flanks of the reference jet and taking difference, we yield

\[
\delta U_+ - \delta U_- = \alpha \left[ U_r \right]_+ - \alpha \left[ U_r \right]_- + (1 + \alpha) \delta y \cdot \left\{ \frac{dU_r}{dy} \right\}_- - \left\{ \frac{dU_r}{dy} \right\}_+ \]

The two latitude boxes are chosen in such a way that the average of the reference wind within the two boxes are equal, therefore,

\[
\delta U_+ - \delta U_- \approx (1 + \alpha) \left\{ \frac{dU_r}{dy} \right\}_- - \left\{ \frac{dU_r}{dy} \right\}_+ \cdot \delta y \quad (A2)
\]

And finally, the shift of the jet is related to the dipolar anomalies and the amplification factor as

\[
\delta y = \frac{\delta U_+ - \delta U_-}{1 + \alpha} \cdot \left\{ \frac{dU_r}{dy} \right\}_- - \left\{ \frac{dU_r}{dy} \right\}_+ \quad ^{-1}
\]

or

\[
\delta y = \frac{U_{rm}}{U_{pm}} \cdot \left( \delta U_+ - \delta U_- \right) \cdot \left\{ \frac{dU_r}{dy} \right\}_- - \left\{ \frac{dU_r}{dy} \right\}_+ \quad ^{-1} \quad (A3)
\]

Appendix B: Moist static stability
The theory of Juckes (2000) and the variant thereof (Frierson 2006; 2008) for the mid-latitude moist static stability relates the bulk moist stability $\Delta \bar{\theta}_e \equiv \bar{\theta}_{et}^* - \bar{\theta}_{es}$, the difference between the tropopause (saturated) and near-surface equivalent potential temperature, to the meridional gradient of surface equivalent potential temperature through a mixing length closure:

$$\bar{\theta}_{et}^* - \bar{\theta}_{es} \sim L \partial_y \bar{\theta}_{es}$$  \hspace{1cm} (B1)

where $L$ can be interpreted as the scale of mid-latitude eddies or the typical width of the storm track. This relation is derived from an argument that the stratification is determined by coupling between moist convection and baroclinic eddies (Juckes, 2000). According to Juckes (2000), moist convection in baroclinic eddies occurs preferentially in the warm sectors of the cyclones, setting the lower limit of the stability. To the extent that moist convection reaches the tropopause and prevents the stratification from being significantly less stable than a moist adiabat, the minimum of the bulk moist stability $\Delta \bar{\theta}_e \equiv \bar{\theta}_{et}^* - \bar{\theta}_{es}$ is approximately 0. Juckes further argues that the additional stabilization above moist adiabat can largely be ascribed to eddy advection of equivalent potential temperature both near the surface and the tropopause, and therefore the mean bulk moist stability should be characterized by the variance of it, $\text{var}(\Delta \bar{\theta}_e)$. The latter can be parameterized in terms of mean meridional gradients through a mixing length closure, giving rise to the relation (B1). This relation may be more intuitively described from a Lagrangian point of view: in a mid-latitude cyclones, the air mass that convects to the tropopause and sets the value of the tropopause moist static energy (and hence the local moist static stability) has its origin some distance equatorward ($\bar{\theta}_{et}^* = \bar{\theta}_{es}^{eq}$) via poleward advection of moist and warm air in warm fronts (Pauluis et al., 2008; Pauluis et al., 2010). $\bar{\theta}_{es}^{eq}$ is usually larger than $\bar{\theta}_{es}$ because for the mean condition, the air masses at lower latitude are usually warmer and moister. The
efficiency of the cyclones in tapping in air with high moist static energy from lower latitudes is proportional to its (i) meridional span (hence the spatial scale $L$ appears); and (ii) meridional gradient of the surface equivalent potential temperature $\Delta_y \theta_{es}$, with larger gradient sustaining more vigorous eddy advection. If one keeps the spatial scale $L$ fixed for different SST boundary conditions, as practiced in predicting the static stability in this study, this is tantamount to fixing the (averaged) moist isentropic slope $L \sim \Delta_z \bar{\theta}_e / \Delta_y \bar{\theta}_{es}$ over the eddy-dominant latitudes. The following schematic illustrates the relationship between parameter $L$ and the moist isentropic slope.

**Figure B1** Schematic for the thermal stratification over the areas dominated by convective baroclinic eddy adjustment. $\Delta_z \bar{\theta}_e \equiv \bar{\theta}_e^t - \bar{\theta}_{es}$ is moist static stability, defined as the difference between the tropopause (saturated) and near-surface equivalent potential temperature. $\Delta_y \bar{\theta}_{es}$ is the difference in near-surface equivalent potential temperature between the in-situ and origin latitudes. See text for detailed description.
References


Atmos. Sci., 47, 2953-2968.


Figure Captions

Figure 1 Upper: latitude of EDSW simulated in (a) the idealized and (b) the full GCMs, respectively. In (b), the crosses mark the cases of double-jet condition. Lower: the latitude of the EDSW versus that of the axis of the storm track in (c) the idealized and (d) the full GCMs. The storm track is measured as the vertically integrated EKE. The dots In (c) and (d) are color coded based on the corresponding $T_m$.

Figure 2 The profiles of 400 hPa zonal wind in the simulations of the full GCM for the cases of $T_m=15\,^\circ$C, $\Delta T$ varying from 10 to 50K at increments of 5K. The cases identified to be in the double-jet regime in Figure 1b are highlighted by the thick lines.

Figure 3 Profiles of $\mathcal{F}_2$ (red), $\mathcal{H}_2$ (dashed red), $\mathcal{F}_1$ (blue), $\mathcal{M}_1$ (dashed back), $U_s$ (black). All the eddy flux terms are normalized by the maximum of $\mathcal{F}_2$ and $U_s$ is normalized by its own maximum.

Figure 4 (left) latitude of max $\xi$ versus that of max EKE; (right) latitude of max $\xi$ versus that of max surface wind in the idealized model.

Figure 5 (a) Baroclinicity $\xi$ (in units of m$^{-1}$s$^{-1}$). Each curve is color coded corresponding to the value of $T_m$ as in Figure 3; (b) Same as (a) but for bulk static stability. Unit = deg Kelvin; (c) Same as (a) but for tropospheric temperature gradients. The black curve represents the gradient of the SST. Unit = K m$^{-1}$.

Figure 6 $\frac{\delta^+ \xi - \delta^- \xi}{\xi_{pm}}$ estimated from model-simulated temperature versus the actual axis of the storm track in the idealized GCM.

Figure 7 Bulk static stability (in deg Kelvin). Theoretical estimations versus simulations by the idealized GCM.
Figure 8 - $\frac{\delta\xi_t}{\xi_{pm}}$ versus the actual displacement of the EDSW in the idealized GCM.

Figure 9 - $\frac{\delta\xi_t}{\xi_{pm}}$ (left) versus the displacement of EDSW axes in the full GCM for all cases; (right) same as (left) but only for the cases of $\Delta T \geq 30^\circ C$. The blue, green, yellow, and red colors represent $T_m = 5^\circ, 10, 15, \text{ and } 20^\circ C$, respectively.
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![Graphs showing EDSW simulations](image)

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