The role of stationary eddies in shaping midlatitude storm tracks

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ABSTRACT

Transient and stationary eddies shape the extratropical climate through their transport of heat, moisture, and momentum. In the zonal mean, the transports by transient eddies dominate over those by stationary eddies, but this is not necessarily the case locally. In particular in storm track entrance and exit regions during winter, stationary eddies and their interactions with the mean flow dominate the atmospheric energy transport. Here it is shown that stationary eddies can shape storm tracks and control where they terminate by modifying local baroclinicity. Simulations with an idealized aquaplanet GCM show that zonally localized surface heating alone (e.g., ocean heat flux convergence) gives rise to storm tracks, which have a well defined length scale that is similar to that of Earth’s storm tracks. The storm tracks terminate downstream of the surface heating even in the absence of continents, at a distance controlled by the stationary Rossby wave length scale. Stationary eddies play a dual role: within half a Rossby wave length downstream of the heating region, stationary eddy energy fluxes increase the baroclinicity and therefore contribute to energizing the storm track; farther downstream, enhanced poleward and upward energy transport by stationary eddies reduces the baroclinicity by reducing the meridional temperature gradients and enhancing the static stability. Transports both of sensible and latent heat (water vapor) play important roles in determining where storm tracks terminate.

1. Introduction

Fluxes of heat, momentum, and water vapor shape Earth’s temperature, wind, and net precipitation distribution. In the extratropics in the zonal mean, these fluxes are dominated by transient eddies (Peixoto and Oort 1992). However, the zonal mean masks the fact that the transient eddies in the Northern Hemisphere are organized into zonally confined storm tracks—regions of enhanced transient eddy kinetic energy (Blackmon 1976) that owe their existence to the zonal asymmetries created by continents (Fig. 1a). (Transient eddies in the Southern Hemisphere are less zonally organized because the zonal asymmetries created by continents are weaker.) To understand the extratropical climate locally, it is therefore essential to understand how the organization of transient eddies into storm tracks is achieved. While it is clear that at least the wintertime storm tracks over the Pacific and Atlantic oceans are generated by the enhanced baroclinicity and heat fluxes from the surface into the atmosphere in the vicinity of the western boundary currents (Fig. 1b, c), it is not clear what controls where the storm tracks terminate. The question of what controls the length scale of storm tracks is the subject of this paper.

Various hypotheses have been proposed to account for the length of storm tracks. Increased surface damping over land has been posited as a reason for reduced transient eddy kinetic energy (EKE) and the termination of storm tracks (Chang and Orlanski 1993; Zurita-Gotor and Chang 2005; Mak and Deng 2007). Hoskins and Valdes (1990) argue that storm tracks may self-destruct altogether if diabatic heating would not enhance them in weakly baroclinic regions, because enhanced transient energy fluxes by the storm tracks reduce baroclinicity. Alternatively, it has been suggested that storm tracks are maintained because of downstream development of eddies that form over the strongly baroclinic storm track entrance regions, and storm tracks then terminate downstream due to either dissipation or barotropic decay of the eddies (Simmons and Hoskins 1980; Orlanski and Katzfey 1991; Chang 1993; Chang and Orlanski 1993; Orlanski 1998; Frisius et al. 1998).

An alternative hypothesis is that orographic stationary eddies play a role in shaping storm tracks of transient eddies. Several studies have indicated that there are interactions between stationary and transient eddies. On the one hand, it has been shown that the dynamic heating associated with stationary eddies and their zonal organization into storm tracks is an important driver of stationary eddies (e.g., Held et al. 2002). On the other hand, GCM
Fig. 1. National Center for Environmental Prediction (NCEP) reanalysis fields averaged over Northern Hemisphere winter (December–February) and over the years 1970–2009. (a) Vertically integrated kinetic energy of 3-10 day bandpass filtered transient eddies (MJ m$^{-2}$). Eddy kinetic energy is defined as $\frac{1}{2} \int (u'^2 + v'^2) \, dp/g$. (b) Absolute value of meridional temperature gradient vertically averaged in the troposphere (K km$^{-1}$). (c) Total upward energy flux from the surface into the atmosphere (W m$^{-2}$) (sum of the latent heat flux, sensible heat flux, and net upward radiative energy flux). (d) 300-hPa geopotential surface anomaly (m). (e) Vertically integrated atmospheric heat flux convergence by transient eddies ($-\nabla \cdot (v'\theta')$, W m$^{-2}$). (f) Vertically integrated heat flux convergence by stationary eddies ($-\nabla \cdot (v^\dagger \theta^\dagger)$, W m$^{-2}$, see definitions in section 3). The color scales for (e) and (f) are logarithmic, with factors of 2 separating contour levels.
simulations have shown that if orography and with it the stationary eddies it generates are reduced, the Northern Hemisphere storm tracks become more zonally symmetric even in the presence of land-ocean contrast (Broccoli and Manabe 1992). Transient eddies in storm tracks thus drive stationary eddies, and stationary eddies (Fig. 1d) in turn shape the storm track (Branstator 1995; Chang et al. 2002; Inatsu et al. 2003). Stationary eddies may thus shape storm tracks, for example, through modifications of the baroclinicity (Lee 1995) or of the barotropic background flow (Manabe and Terpstra 1974; Lee 1995; Harnik and Chang 2004; Son et al. 2009; Park et al. 2010; Brayshaw et al. 2009; Saulière et al. 2011).

How stationary eddies may lead to the termination of storm tracks through baroclinic processes has received less attention in the recent literature, possibly because in the zonal mean, transient eddies dominate the poleward and upward transport of sensible and latent heat. Locally, however, energy fluxes associated with stationary eddies can be significant (Lau 1979a,b) and, in fact, they dominate those associated with transient eddies (Figs. 1e, f). The stationary eddy fluxes include components owing to eddy–mean flow interactions, namely, owing to the advection of the zonal-mean moist static energy (MSE) by stationary eddies and to the advection of the stationary-eddy MSE by the zonal-mean flow. (The relative contribution of the various parts of the MSE are discussed in greater detail in section 3.) The energy flux divergences are locally dominated by stationary eddies. It is plausible, then, that baroclinic modifications of the background flow by stationary eddies exert a substantial influence on storm track organization.

Here we use simulations with an idealized aquaplanet GCM to study systematically how transient and stationary eddies interact and shape storm tracks. In the GCM, we generate storm tracks by prescribing a zonally localized surface heating. Similar to those in Earth’s atmosphere, the so generated storm tracks have a well defined extent, whose dynamical origin we investigate by varying the planetary rotation rate. We use a GCM with a representation of the atmospheric hydrologic cycle to capture moist processes that are suspected to be important for storm track dynamics. For example, the generation of eddy energy through the release of latent heat in phase changes of water in warm sectors of cyclones can energize transient eddies and can contribute to the maintenance of storm tracks (Hoskins and Valdes 1990; Chang et al. 2002). At the same time, the release of latent heat can stabilize the thermal stratification, and meridional latent heat transport can reduce meridional temperature gradients, thus reducing baroclinicity and potentially damping transient eddies (e.g., Lee and Held 1993; Schneider and O’Gorman 2008). The relative importance of these competing roles of latent heat release—energizing eddies and reducing baroclinicity—is insufficiently understood (Schneider et al. 2010). Both may play a role in shaping storm tracks.

The model is described in section 2. In section 3, the stationary and transient eddies generated in response to a localized surface heating are analyzed and are shown to resemble those seen in Earth’s atmosphere. Section 4 presents simulations with different planetary rotation rates, which allow us to study storm tracks whose length varies from a fraction to the full length of an extratropical latitude circle. We find two key stationary-eddy mechanisms that control where storm tracks terminate. In the first, stationary eddies increase meridional temperature gradients for half a Rossby wave length downstream, mainly because of zonal energy fluxes by stationary eddies. In the second, stationary eddies reduce baroclinicity farther downstream because they reduce meridional temperature gradients and enhance the static stability through their poleward and upward transport particularly of latent heat. These mechanisms are analyzed in section 5.

2. Idealized GCM with zonal asymmetries

a. Model

We use an idealized GCM with a simple representation of moisture. The model is based on the GFDL Flexible Modeling System (GFDL 2004). It is a spherical-coordinate primitive-equation model of an ideal-gas atmosphere, which is similar to that used by O’Gorman and Schneider (2008b) and Frierson et al. (2006). For our reference simulation, we use a horizontal resolution of T42 and 30 vertical sigma levels. For simulations with higher planetary rotation rates, we use the higher horizontal resolutions of T85 and T127, to resolve the smaller eddies (see Table 1). To damp small scales, scale-selective eighth-order hyperdiffusion is included in the vorticity, divergence, and temperature equations.

The lower model boundary is a uniform slab ocean with no dynamics but with local energy balance. The surface temperature is not prescribed, unlike in many previous studies of storm tracks (e.g., Inatsu et al. 2002, 2003; Brayshaw et al. 2008, 2009, 2011; Saulière et al. 2011), but it changes in response to changing surface fluxes of radiative energy, sensible heat, and latent heat. A planetary boundary layer scheme with Monin-Obukhov surface fluxes, which depend on the stability of the boundary layer, links atmospheric dynamics to surface fluxes of momentum, latent heat, and sensible heat. Radiative transfer is represented by a two-stream gray radiation scheme with longwave and shortwave optical depths that only depend on latitude and pressure, and a top-of-the-atmosphere insololation which is approximately equal to Earth’s annual-mean insololation. Moist convection is represented by a slightly modified version (O’Gorman and Schneider 2008b) of the quasi-equilibrium convection scheme of Frierson (2007). It relaxes temperature profiles toward a moist-adiabatic pro-
Table 1. The rotation rate, resolution, and resulting energy-containing wavenumber of the simulations. All other parameters are the same in all cases.

<table>
<thead>
<tr>
<th>Rotation rate (multiples of (\Omega_e))</th>
<th>Resolution</th>
<th>Energy-containing wavenumber</th>
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<tr>
<td>12</td>
<td>T127</td>
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<td>8</td>
<td>T127</td>
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<td>T85</td>
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<td>T42</td>
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<td>1/4</td>
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To study storm tracks in an aquaplanet idealized GCM, we must introduce some zonal asymmetries. Such asymmetries can be introduced in several ways, for example, by using orography (e.g., Son et al. 2009; Brayshaw et al. 2009), or by thermal forcing (e.g., Chang 2009; Brayshaw et al. 2011). In this study, we do so by applying a localized heat flux convergence (a “Q-flux”) at the model’s bottom boundary slab ocean. This creates a localized heat source, representing a warm area in the ocean such as the Gulf Stream or Kuroshio western boundary currents. We apply the localized surface heating over a triangular area in the northern hemisphere between latitudes 25°N and 50°N (see Fig. 2), crudely representing the shape of the western oceanic boundaries (both North America and Asia have an eastern boundary with a coast line extending in the south-west to north-east direction). The particular shape, size, and magnitude of the local heat source do not affect our results qualitatively. To allow comparison with a control case, we add in the southern hemisphere the same ocean heat flux convergence but spread over all longitudes in a zonally symmetric way. This allows us to treat the southern hemisphere a control case with no zonal asymmetry and with the same zonal-mean thermal driving.

The magnitude of the localized surface heating is chosen so that the enhanced temperature gradients near the forcing region are similar to those observed in the entrance region of the northern hemisphere storm tracks in winter (compare Fig. 2b and Fig. 1b). We achieve temperature gradients with peak values near \(1.3 \times 10^{-5} \text{ K km}^{-1}\) with...
a Q-flux of 850 W m$^{-2}$. We will refer to this case as our reference simulation, which provides us with the reference climate that we study here. Experiments with different magnitudes of the Q-flux will be presented in a forthcoming paper; the magnitude of the Q-flux is not essential for the results on which we focus here.

To allow a wide variation of length scales of both the eddies and the storm tracks we vary the rotation rate of the planet. We use simulations with rotation rates between 1/4 and 12 times that of Earth. A summary of the experiments, the model resolutions, and the resulting energy-containing wavenumbers (defined in section 4) is given in Table 1. All experiments are spun-up to for at least 2000 days to a statistically steady state. The time-mean results presented here are averaged for at least a subsequent 700 days.

c. Response to a localized heat source

In the reference simulation, the localized surface heating results in a localized downstream (to the east) increase in transient EKE in a well defined storm track (Fig. 2a). Similar storm track generation was seen in previous studies with prescribed sea surface temperatures (Inatsu et al. 2002). EKE has a peak downstream of the perturbation region and then decays farther downstream to values lower than in the control southern hemisphere. Thus, the existence of a region of localized surface heating creates not only a region of high EKE immediately downstream, due to enhanced baroclinicity and diabatic heating, but also a region of damped EKE farther downstream, both compared with the southern hemisphere and with a case with no additional (zonally asymmetric or symmetric) heating (Kaspi and Schneider 2011a). The surface temperature is increased in the heating region and immediately downstream of it owing to advection by the mean flow. In addition, surface temperatures are lowered immediately upstream of the heating region. This is an upstream response set by stationary waves, which generate a negative temperature anomaly within a Rossby wave plume (Kaspi and Schneider 2011b). Furthermore, the localized surface heating leads to an increase in the meridional temperature gradient poleward of the heating region and immediately downstream of it (Fig. 2b).

Comparing Fig. 2a, b to Fig. 1a, b shows that EKE and meridional temperature gradients in our reference simulation match the reanalysis data well. The storm track EKE peaks downstream and poleward of the source region and is elevated over a similar distance downstream as on Earth, although the idealized model has no continents or topography.

![Energy flux convergence](image-url)
3. Interaction between stationary and transient eddies

The response of the system to the localized surface heating consists of a locally enhanced transient EKE (the storm track) and the formation of stationary eddies. To understand the relative contribution of transient and stationary eddies to the energy budget, we compare simulations with a localized zonally asymmetric surface heating with simulations without zonal asymmetries at the boundaries. Without zonal asymmetries, the stationary eddy contribution vanishes as expected (Fig. 3a). The transient eddy fluxes of dry static energy (DSE), \( \cdot \), and latent energy (LE), \( \cdot \), both diverge in the tropics and subtropics and converge in the extratropics. For DSE and LE we use standard definitions (Peixoto and Oort 1992), with the specific heat of dry air \( c_p = 1004 \text{ J kg}^{-1} \text{ K}^{-1} \) and the latent heat of vaporization \( L = 2.5 \times 10^6 \text{ J kg}^{-1} \), consistent with the thermodynamics of our GCM. Thus, as is well known, the poleward energy flux\(^1\) owing to transient eddies reduces the pole-to-equator temperature gradients. But with zonal asymmetries at the boundaries, the picture becomes more complicated: stationary eddies contribute significantly to the energy flux convergence (Fig. 3b).

To define stationary and transient eddy fluxes locally, we split fields into deviations from the time and zonal means. We denote by \( \cdot \) the deviation from the time mean (\( \cdot \)), so that \( \cdot = \cdot + \cdot \). Then, for example, the time mean meridional DSE flux can be split so that

\[
\overline{\overline{s}} = \overline{\overline{s'}} + \overline{\overline{s"}} + \overline{[\overline{\overline{\overline{s'}}}]} + \overline{[\overline{\overline{s"}}]} . \quad (1)
\]

Zonally averaged, the third and fourth terms on the right-hand side vanish, leaving just the contribution from the transient eddies, the stationary eddies, and the mean meridional advection (e.g., Peixoto and Oort 1992),

\[
[\overline{\overline{s}}] = [\overline{\overline{s'}}] + [\overline{\overline{s"}}] + [\overline{\overline{\overline{s'}}}] . \quad (2)
\]

Locally, however, the cross terms in (1) are significant, representing stationary eddy-mean flow interactions, so that the stationary eddy response locally contains the second, third, and fourth term in (1). We define the total stationary eddy response due to the existence of zonal asymmetries as

\[
\overline{\overline{\overline{s'}}} = \overline{\overline{s}} - [\overline{\overline{s}}] = \overline{[\overline{\overline{s'}}]} + \overline{[\overline{s"}]}, \quad (3)
\]

which includes the interaction with the mean flow. An analogous analysis applies to the zonal and vertical components of the energy flux.

\(^1\)When we refer to the energy flux, we refer to the static component of the energy flux.

To analyze how the stationary and transient eddy energy flux terms affect temperature gradients and baroclinicity locally, it is useful to decompose the energy budget into the terms in (1) so that

\[
\frac{\partial \overline{s}}{\partial t} = \nabla \cdot (\overline{\overline{\overline{s'}}}) - \nabla \cdot (\overline{\overline{\overline{s"}}}) - \\
\nabla \cdot ([\overline{\overline{\overline{s'}}}] + \nabla \cdot (\overline{\overline{\overline{s"}}})) - \nabla \cdot ([\overline{\overline{\overline{s'}}}] + \overline{\overline{\overline{s"}}}) + Q, \quad (4)
\]

where \( Q \) is the diabatic heating. The time derivative vanishes in a statistically steady state. On the right-hand side, there is the flux convergence of the transient eddies (first term), the flux convergence of the stationary eddies (second term), advection of the stationary eddy DSE by the zonal-mean flow (third term), advection of the zonal-mean DSE by the stationary eddies (fourth term), and advection of the zonal-mean DSE by the zonal-mean flow (fifth term). In the zonal mean, \([\overline{\overline{\overline{s'}}}] = \overline{\overline{\overline{s"}}}] \), the contribution from the stationary eddies in the northern hemisphere (where the localized surface heating was added) is of the same magnitude as the contribution of the transients, both for the DSE and LE flux convergence (Fig. 3b), as is the case in Earth’s atmosphere (Peixoto and Oort 1992). In the southern hemisphere (where the added surface heating is zonally symmetric), the stationary eddy contribution vanishes (up to sampling error). The southern hemisphere transient eddy contribution in the reference simulation (Fig. 3b) is not identical to that in the simulation without added surface heating (Fig. 3a) because the additional, zonally symmetric, surface heating broadens the flux convergence poleward.

Locally, however, the contribution of the stationary eddies is much larger than that of the transients. Figure 4 shows the vertically integrated transient eddy (top) and stationary eddy (bottom) DSE (left) and LE (right) flux convergences. The stationary eddy DSE flux convergence is locally the dominant term in the northern hemisphere, particularly near the region of the localized surface heating (Fig. 4c). Upstream of the localized heating, there is strong divergence, indicating dynamic cooling. The most significant feature is that downstream of the localized heating, there is a strong upgradient energy flux (relative to the mean pole-to-equator gradient), which is evident by the strong front between convergence to the south and divergence to the north of the storm track. This energy flux opposes the tendency of the transient eddies to reduce pole-to-equator temperature gradients (Fig. 4a). Therefore, one reason storm tracks do not self-destruct immediately downstream of their entrance regions (Hoskins and Valdes 1990) is that stationary eddies (including their interaction with the mean flow) act against the transients to maintain baroclinicity. They do so primarily through the zonal advection of DSE by the mean flow, and over a length scale controlled by stationary eddies (see discussion below). Farther downstream, the effect of stationary...
Fig. 4. Vertically integrated convergence, of DSE and LE fluxes. (a) Transient eddy DSE flux convergence. (b) Transient eddy LE flux convergence. (c) Stationary eddy DSE flux convergence. (d) Stationary eddy LE flux convergence. These are the same fields as in Fig. 3 but shown only temporally (not zonally) averaged. The color scale is logarithmic, with factors of 2 between contour levels.

Eddies on the pole-to-equator temperature gradients is reversed: weaker convergence or divergence to the south, and convergence to the north of the storm track indicate that stationary eddies reduce baroclinicity there. This causes the self-destruction of EKE downstream of the storm track (Kaspi and Schneider 2011a).

The most significant feature of the stationary LE flux (Fig. 4d) is the strong convergence in midlatitudes, on the equatorward flank of the storm track, and the strong divergence over and equatorward of the surface heating region. Thus, there is a LE flux from the tropics equatorward of the heating region into midlatitudes. This is similar to the “moisture rivers” seen in observations (e.g., Newell et al. 1992; Zhu and Newell 1998), which carry out much of the poleward moisture transport in Earth’s atmosphere. This moisture transport occurs in filaments extending off extratropical cyclones toward the equator; hence, the moisture rivers are often associated with transient eddies. However, because the filaments occur at particular locations—for example, extending from warm oceanic regions such as the Pacific Warm Pool and the Gulf of Mexico into the storm tracks, or, in our simulations extending from south of the localized heating—they appear as part of the stationary moisture flux term $\nabla \cdot (\bar{u} \bar{q})$. Similar LE stationary transport has been also associated with subtropical anticyclones (Shaw and Pauluis 2012). The moisture transported into midlatitudes by stationary eddies is then picked up by the transient eddies (Fig. 4b) and transported farther poleward, as indicated by the pattern of divergence and convergence of the transient LE flux in the extratropics (Fig. 4b).

To understand the role of the various components of the stationary eddy fluxes, we further split the total stationary eddy energy flux (Fig. 4) into the individual components (Eq. 4). This includes the terms owing to interactions with the mean flow. The pure stationary eddy terms (Fig. 5a, b) are smaller in magnitude than the terms involving the zonal-mean flow (Fig. 5c–f), although the latter vanish in the zonal mean. Downstream of the surface heating region, both the DSE (Fig. 5a) and LE (Fig. 5b) components of the pure stationary flux terms are similar in magnitude, but opposite in sign, to the transient energy flux terms (Fig. 4a, b). However, these terms do not completely compensate each other either locally or in the zonal mean. The main contribution to the stationary eddy DSE flux convergence $-\nabla \cdot (\bar{u} \bar{s})$, is from the advection of the stationary eddy DSE by the zonal-mean flow (third term on the right-hand side of Eq. 4, Fig. 5c). To leading order, this term is dominated by the zonal flow $\nabla \cdot (\bar{u} \bar{s}) \approx |\bar{u}| \partial_x \bar{s}$, where $\partial_x$ denotes the longitudinal derivative. Thus, the upgradient energy flux relative to the mean equator-to-pole temperature gradient that acts against the transient downgradient flux (Fig. 4a), is caused primarily by zonal downstream energy fluxes, which extend approximately over half a Rossby wave length (see section 5). The wave motion of the stationary eddies (creating warm and cold regions
Fig. 5. Components of the stationary eddy energy flux convergence (vertically integrated). (a) Stationary eddy DSE flux convergence, $-\nabla \cdot (\mathbf{u}^s_\ast \mathbf{s}^\ast)$. (b) Stationary eddy LE flux convergence, $-L \nabla \cdot (\mathbf{u}^q_\ast \mathbf{q}^\ast)$. (c) Advection of stationary eddy DSE by the mean flow, $-\nabla \cdot (|\mathbf{u}|\mathbf{s}^\ast)$. (d) Advection of stationary eddy LE by the mean flow, $-L \nabla \cdot (|\mathbf{u}|\mathbf{q}^\ast)$. (e) Advection of the zonal-mean DSE by stationary eddies, $-\nabla \cdot (\mathbf{u}^s |\mathbf{s}|)$. (f) Advection of the zonal mean LE by stationary eddies, $-L \nabla \cdot (\mathbf{u}^q |\mathbf{q}|)$. The color scale is logarithmic, with factors of 2 between contour levels.
downstream) controls the length scale over which the downstream advection occurs, and therefore controls where the stationary energy flux maintains the mean equator-to-pole temperature gradient and where it reduces it.

The stationary LE flux convergence has contributions both from the pure stationary eddies (Fig. 5b), the zonal-mean advection of the stationary fluctuations (Fig. 5d), and from the advection of the mean LE by the stationary eddies (Fig. 5f). The sum of these three contributions (Fig. 4d) is comparable in magnitude and opposite in sign to the transient LE flux convergence (Fig. 4b). The moisture source regions of the stationary and transient LE fluxes differ: the stationary fluxes originate primarily in low latitudes, whereas the transient fluxes originate in mid-latitudes. However, both are manifestations of moisture fluxes in filamentary extensions of extratropical cyclones. Comparing the relative contributions of DSE and LE to the total energy flux shows the importance of LE fluxes in maintaining the storm track.

4. Effect of rotation

The map of vertically integrated EKE (Fig. 1a) shows that the northern hemisphere storm track EKE weakens over land. Models with continents show a similar behavior (e.g., Chang 2009), often attributed to larger dissipation over the continents (e.g., Chang et al. 2002; Mak and Deng 2007), barotropic decay (Simmons and Hoskins 1980), or influences of orographic stationary waves (Saulièrie et al. 2011). However, in our simulations even without any continents or varying surface friction, the storm track terminates over roughly the same length scale as in observations (Fig. 2a). In this section, we investigate the mechanisms controlling the termination of the storm track, using a series of simulations in which the rotation rate of the planet is varied. Using this approach has several advantages. First, since the storm track length in the model is close to that observed on Earth, changing the rotation rate allows creating a larger scale separation between the length of the storm track and the circumference of the planet, thus giving more room to study the mechanisms controlling the storm track length. Second, varying the rotation rate may allow separating the relative importance of the stationary wave length scale and the transient eddy length scale. Stationary ($\omega = 0$) barotropic Rossby waves in a constant background flow have the dispersion relation

$$0 = \omega = \bar{u}k - \frac{\beta k}{K^2}, \quad (5)$$

where $\beta$ is the planetary vorticity gradient, $K^2 = k^2 + l^2$, and $k$ and $l$ are the zonal and meridional wavenumbers (Pedlosky 1987). Therefore, the stationary wave length scale

$$L_s = 2\pi \left(\frac{\pi}{\beta}\right)^{1/2}, \quad (6)$$

Fig. 7. EKE (MJ m$^{-2}$) in the northern hemisphere (red) and southern hemisphere (blue) of the simulations with added heating, and in a control simulation without added heating (black), all vertically integrated and averaged over the baroclinic zone in each hemisphere. The different panels correspond to different planetary rotation rates (in multiples of Earth’s rotation rate $\Omega_e$). Dashed gray lines mark the region of the localized surface heating in the northern hemisphere.
Fig. 6. (a) Vertically integrated tropospheric EKE for planetary rotation rates of $0.5 \Omega_e$, $2 \Omega_e$, $4 \Omega_e$, and $8 \Omega_e$. (b) Corresponding vertically integrated stationary meridional MSE flux ($\bar{\pi} \bar{m}^l$). The units are non-dimensional, with 1 corresponding to 1, 0.6, 0.25, and 0.15 MW m$^{-1}$ for planetary rotation rates of $0.5 \Omega_e$, $2 \Omega_e$, $4 \Omega_e$ and $8 \Omega_e$, respectively. The EKE in a similar simulation with planetary rotation rate $\Omega = \Omega_e$ is shown in Fig. 2.

varies as the inverse of the square root of the rotation rate if $\pi$ variations can be neglected. By contrast, measures of the transient eddy length scale such as the deformation radius vary inversely with rotation rate if variations in the static stability can be neglected (Schneider and Walker 2006; Merlis and Schneider 2009). Comparing the transient EKE for four cases with the same thermal driving but rotation rates of 1/2, 2, 4, and 8 times that of Earth ($\Omega_e$) shows that the faster the rotation rate, the more zonally localized the storm track becomes (Fig. 6a). Locally in the storm track region EKE is enhanced, however, consistently for all cases, away from the localized heating, the EKE is weaker than the EKE in the reference southern hemisphere (Kaspi and Schneider 2011a). As the rotation rate increases, EKE in the zonally symmetric southern hemisphere decreases because the baroclinic length scale decreases, so that the latitudinal region over which baroclinic instability occurs becomes more confined, thus decreasing the efficiency with which eddies convert mean available potential energy (MAPE) to EKE. This happens even though the smaller eddy length scales cause a reduction in the overall poleward heat flux so that the overall equator-to-pole temperature gradient is larger (Schneider and Walker 2008). Figure 6b shows the corresponding stationary meridional flux ($\bar{\pi} \bar{m}^l$). As expected from Eq. (6), the stationary eddy length scale decreases with increasing rotation rate.

To compare the downstream evolution of the storm track in the northern hemisphere with the reference southern hemisphere, Figure 7 shows the EKE in each hemisphere averaged over the baroclinic zone defined as the latitudinal region between latitude 30° and where the EKE reduces to 80% of its southern hemisphere zonal mean maximum value. The two main features are that the EKE increases sharply immediately downstream of the localized heating in the northern hemisphere (red line), and that farther downstream the storm track self-destructs so that the EKE in the northern hemisphere is smaller than that in the southern hemisphere (blue line). Moreover, the EKE in the northern hemisphere is still weaker than in a simulation with no added heating at all (black line), meaning that although the existence of a localized heating causes a local increase in EKE, it also creates a stationary eddy response that causes EKE self-destruction farther downstream (Kaspi and Schneider 2011a). EKE is reduced downstream because of stationary MSE fluxes, which we discuss in detail in section 5. To get a consistent measure of the length of the storm track, we define the length of the storm track ($L_{ST}$) as the distance from the localized heating to the point downstream at which the northern hemisphere EKE is equal to the zonal-mean southern hemisphere EKE.

For different rotation rates, the length of the storm track scales nearly as $\Omega^{-1/2}$ (Fig. 8). It also scales fairly well with the length scale of the stationary wave (Eq. 6), using for $\pi$ the maximum value of the vertically and zonally averaged zonal wind. (This value of $\pi$ varies by roughly a factor of four over this range of rotation rates.)

In dry models, transient eddy length scales are similar to the stationary wave length scale and the transient eddy length scale other factors beside the rotation rate (e.g., mean wind velocity and static stability) are important; however, we find that for the range of rotation rates we use (factor 48 variations), these factors vary on average by less than a factor of four, and therefore varying the rotation rate is a good method for varying these length scales.

2Both for the stationary wave length scale and the transient eddy length scale other factors beside the rotation rate (e.g., mean wind velocity and static stability) are important; however, we find that for the range of rotation rates we use (factor 48 variations), these factors vary on average by less than a factor of four, and therefore varying the rotation rate is a good method for varying these length scales.

3This relation breaks for slower rotation rates, for which the length of the storm track approaches the length of a latitude circle.
to the Rossby deformation radius (Schneider and Walker 2008; Merlis and Schneider 2009). However, when including the effects of moisture, it is more difficult to represent the eddy length scale with simple scalings. To estimate the transient eddy length scale in the simulations without the localized heating, we use the energy-containing wavenumber $n_e$ defined by

$$n_e (n_e + 1) = \sum E_n \sum [n (n + 1)]^{-1} E_n,$$

where $n$ is the spherical wavenumber, and $E_n$ is the EKE spectrum omitting the zonal wavenumber zero, thus omitting energy from the zonal-mean flow (Schneider and Walker 2008). We then define the energy-containing eddy length scale to be

$$L_e = \frac{2\pi a}{\sqrt{n_e (n_e + 1)}},$$

where $a$ is Earth’s radius. Overlaying this energy-containing length scale with the storm track and stationary waves length scales shows that the energy-containing length scale is very similar to the stationary wave scale (Fig. 8). This similarity in length scales arises because even in moist atmospheres EKE appears to scale linearly with dry MAPE (O’Gorman and Schneider 2008a), implying there is a linear relation between the Rhines scale

$$L_\beta = \frac{2\pi (u'/\beta)^{1/2}},$$

and the stationary wave length (Eq. 6). Thus it is difficult to separate the stationary and transient length scales.

5. Mechanisms linking stationary and transient eddies

a. Modification of temperature gradients

Section 3 showed the separate contributions of dry and moist processes to the energy flux, where in both cases the energy flux is locally dominated by stationary eddies. In this section, instead of focusing on the individual contributions of dry and moist processes, we look at the combination of the two and focus on the MSE, $m = c_p T + g z + L q$. This allows us to concentrate on how the total dynamic heating affects temperature gradients and baroclinicity. As in section 3, we begin with presenting zonally averaged fields. The nonlinear interaction between the transient and stationary eddies can be seen already in the zonally symmetric fields (for which $[\overline{\nu' \overline{m'}}] = [\overline{\nu' \overline{m'}}]$).

In Fig. 9, the zonally averaged poleward MSE flux $\overline{\nu' m'}$ is decomposed into the transient eddy component $\nu' \overline{m'}$ and the stationary eddy component $\overline{\nu' m'}$. For the case with

\[ A \text{ linear relation between MAPE and EKE implies that the Rossby radius } L_R \text{ and the Rhines scale } L_\beta \text{ (the expected energy-containing scale) are linearly related, } L_\beta \sim L_R. \text{ (Schneider and Walker 2006). Using this result, the scaling relations in Schneider and Walker (2006) and thermal wind balance then imply } u' \sim EKE^{1/2} \sim \overline{\nu}. \]
no zonal asymmetries (black lines), the stationary eddy component is zero. The total energy transport reaches 4-5 PW, similar to observations for the southern hemisphere during winter (Peixoto and Oort 1992). As is well known (e.g., Vallis 2006), the transient eddy component has larger amplitude than the total in the extratropics because the mean-flow component is negative (due to the thermally indirect Ferrel cell). For the zonally asymmetric case with localized heating (red lines), in the southern hemisphere where the additional heating is spread out zonally symmetrically, both the total energy flux and the transient eddy energy flux are strengthened compared with the unperturbed case. However, in the northern hemisphere where the additional heating is zonally localized, the transient eddy energy fluxes are weakened. The stationary eddy energy fluxes strengthen and overcompensate this weakening of the transients, so that the total poleward energy flux strengthens by about 20%. This implies that transient eddies and stationary eddies in storm tracks interact nonlinearly (through their modification of the mean flow) and cannot be considered in isolation of each other. The MSE flux in our idealized model matches that estimated from observations (Oort and Peixóto 1983; Trenberth and Caron 2001). This increase in poleward energy flux as zonal asymmetries are introduced causes the reduction of meridional temperature gradients, resulting in reduced baroclinicity and the termination of the storm track downstream.

The meridional MSE fluxes showed the importance of the stationary eddies for the poleward energy transport,
but to understand the localization of storm tracks, we next look at the vertically integrated MSE flux convergence (Fig. 10). The individual (dry and moist) components of the MSE flux convergence (Fig. 4) showed that the stationary eddy MSE flux convergence locally dominates. Correspondingly dividing the MSE flux convergence (Fig. 10c) into the individual components shows that it is likewise dominated by the stationary components (Fig. 10e), not by the transient component (Fig. 10a). Among the stationary eddy terms, the dominant term is the advection of the stationary eddy MSE by the zonal-mean wind (Fig. 10c), where the zonal component is dominant so that \( \nabla \cdot (\overline{u^m}) \approx \overline{\partial_x u^m} \). Thus the mean zonal advection causes a strong divergence of MSE flux northeast of the localized heating (mainly because of downstream advection of cold air), which increases local temperature gradients downstream of the localized heating (Fig. 13a). This happens over one half Rossby wavelength downstream of the heating region since the temperature fluctuation of the wave, which is zonally advected, controls the pattern of dynamical heating and cooling. This is a main reason for the maintenance of the storm track, for which this process acts against the tendency of the transient MSE flux to reduce temperature gradients and therefore weaken the storm track. A similar pattern is found for other rotation rates (Fig. 11).

To demonstrate how the stationary waves set the length of the storm track, Fig. 12 shows the meridional gradient of the MSE flux convergence, \( -\partial_y \nabla \cdot (\overline{u^m}) \), integrated over the baroclinic zone (blue line). Negative values indicate flux that increases meridional temperature gradients, and positive values a flux that reduces temperature gradients. Over a range of rotation rates, this quantity anti-correlates with the storm track EKE (red line), meaning that the enhancement of the storm track happens where the stationary eddies increase baroclinicity, thus giving a direct link between the stationary eddies and enhanced baroclinicity.

The localized heating induces also a surface cyclone to the east of it (Hoskins and Karoly 1981). This causes warm air advection from lower latitudes, and enhanced dynamical heating east of the localized surface heating for half a stationary wave length scale downstream of the localized heating (Fig. 10d) \(^5\). The combination of this dynamical heating by the stationary wave (Fig. 10d) and the advection of stationary eddy MSE fluctuation by the mean flow (Fig. 10c), which are both controlled by the stationary eddy length scale, determine the length scale over which the storm track terminates. Over the first half stationary wave length scale this pattern enhances temperature gradients and thus the baroclinicity because of dynamical heating to the south and cooling to the north of the storm track (see

\(^5\)The Icelandic and Aleutian lows on Earth are similar phenomena although they presumably arise not only because of local heating (over the Gulf Stream in the Atlantic and Kuroshio in the Pacific), but also because of orography.
Fig. 12. Meridional gradient of the vertically integrated stationary eddy MSE flux convergence (e.g., Fig. 11b) integrated over midlatitudes $-\int_{\phi_1}^{\phi_2} \frac{\partial}{\partial \phi} \left( \nabla \cdot \left( \mathbf{u}^\dagger \mathbf{m}^\dagger \right) \right) \, d\phi$. Here, $\phi_1 = 30^\circ$N and $\phi_2$ is the northern edge of the baroclinic zone, defined as the latitude where the EKE decreases to 80% of the maximum southern hemisphere zonal mean EKE (blue). (For rotation rates $1/2, 1, 4/3, 2, 3, 4, 6, \text{and } 8\Omega_e$, $\phi_2$ is 67, 61, 59, 56, 52, 49, 45, and 43$^\circ$N, respectively.) For comparison, the EKE difference between the northern and southern hemispheres integrated over the same latitudes is shown in red (right axis scale). The individual components of the gradients of the stationary eddy MSE convergence are the dashed lines: advection of the stationary eddy MSE perturbation by the mean flow $-\int_{\phi_1}^{\phi_2} \frac{\partial}{\partial \phi} \left( \mathbf{u} \cdot \nabla \mathbf{m} \right) \, d\phi$ (magenta); zonal-mean stationary eddies $-\int_{\phi_1}^{\phi_2} \frac{\partial}{\partial \phi} \left( \nabla \cdot \left( \mathbf{u} \mathbf{m} \right) \right) \, d\phi$ (green); advection of the mean MSE gradient by stationary eddies $-\int_{\phi_1}^{\phi_2} \frac{\partial}{\partial \phi} \left( \nabla^t \frac{1}{\rho} \frac{\partial}{\partial \phi} \mathbf{m} \right) \, d\phi$ (black). Negative values indicate a tendency to increase meridional temperature gradients. The dashed gray lines indicate the longitudes of the northern hemisphere localized heating.
Fig. 13. (a) Difference in absolute value of meridional temperature gradients between the northern hemisphere and the southern hemisphere zonal mean. (b) Vertically averaged dry static stability ($N^2 = \frac{g}{\theta} \frac{\partial \theta}{\partial z}$) difference between the northern hemisphere and the southern hemisphere zonal mean.

how the combination of Fig. 10c and Fig. 10d lead to Fig. 10e); farther downstream it decreases baroclinicity because the stationary components of the poleward heat flux contribute to reduction of meridional temperature gradients (Fig. 12).

The difference in meridional temperature gradients between the northern hemisphere and the southern hemisphere zonal mean allows us to quantify the effect of the localized heating on the meridional temperature gradients. Figure 13a shows that beyond the region of local increase in the northern hemisphere downstream of the localized heating, the temperature gradient is generally smaller everywhere else around the latitudes of the perturbation. The addition of $\sim$1 PW of poleward energy transport (Fig. 9) results in a decrease in vertically averaged temperature gradients on the order of 0.3 K per 100 km in midlatitudes. This decrease in temperature gradients is a key reason for the decrease in baroclinic conversion downstream, which results in the decrease in EKE (Figs. 6, 7) and the termination of the storm track. For all rotation rates in Fig. 8, there is a $\sim$1 PW increase in poleward MSE flux compared to the zonally symmetric southern hemisphere, and $\sim$2 PW increase compared to a case with no added heating (Kaspi and Schneider 2011a). Similarly, Earth’s northern hemisphere has a poleward MSE flux that is $\sim$2 PW larger than that of the southern hemisphere, where the contribution of stationary eddies is smaller (Peixoto and Oort 1992).

In summary, we have shown that, in the zonal mean, the existence of localized heating and the stationary eddies it generates cause an increase in the zonal-mean poleward MSE flux and therefore a reduction of zonal-mean temperature gradients, resulting in less baroclinicity and less EKE. Over half a stationary wave length downstream of the localized heating, however, baroclinicity is increased mainly by downstream advection of MSE by stationary eddies (Figs. 10, 12). This opposes the tendency of the transient eddies to decrease baroclinicity. Hoskins and Valdes (1990) pointed to the possibility that latent heat release acts in some way to fuel the storm track and oppose the destruction of mean gradients by the transient eddies. Here we have shown that stationary eddies, induced by the localized heating and with contributions from both the DSE and LE components, lead to energy fluxes that enhance baroclinicity immediately downstream of the heating, while the transients reduce baroclinicity.

b. Modification of static stability

MAPE and baroclinic generation are controlled not only by the meridional temperature gradients but also by the static stability (Pedlosky 1987). Therefore, to quantify the baroclinicity in the storm track entrance and exit regions, the effect of the eddies on the static stability should also be considered. Recent studies have shown that the static stability is important for controlling the latitudinal position of the storm track (Lu et al. 2010) and its intensity during climate change (O’Gorman 2010). The dry static stability difference between the northern hemisphere and the southern hemisphere zonal mean in our reference simulation shows that in the storm track and particularly upstream of the entrance region, there are regions of enhanced dry static stability (Fig. 13b)\(^6\). High static stability results in reduced baroclinicity and therefore, in combination with the reduced temperature gradients downstream, results in the localization of the storm track. To understand the mechanisms leading to the increase in static stability, we look separately at the regions upstream and downstream of the localized heating.

Upstream, there is advection of cold polar air near the surface, resulting from the surface cyclone created in response to surface heating (Hoskins and Karoly 1981). The longitudinal length scale of this region is controlled by the stationary Rossby wave plume induced by the localized heating (Rhines 2002; Kaspi and Schneider 2011b). This cold near-surface air decreases the lapse rate near the surface, and therefore increases the static stability. The region of increased static stability upstream correlates with the region of cold temperature anomalies (Kaspi and Schneider\(^6\)).

\(^6\)We use the dry static stability, which we find for these cases not to differ much in the extratropics from moist measures such as that of O’Gorman (2011).
2011b). Similar cold regions with high static stability near the surface (and even inversions) are found near the eastern continental boundaries during winter (Lee and Mak 1994).

**Downstream**

The increase in static stability is mainly because of LE release. The stationary moisture fluxes bring streams of moisture from the tropics to the extratropics (Fig. 4d), and this moisture is then picked up by the transient eddies and carried farther northward. The air cools adiabatically as it moves upward and poleward (Fig. 14b), resulting in condensation and latent heat release aloft in the mid-troposphere (Fig. 15b). This heating reduces the lapse rate (Fig. 14c) and increases the static stability (Fig. 14d), resulting in reduced baroclinicity. As seen horizontally in Fig. 5c and vertically in Fig. 15e, the zonal-mean advection of the stationary eddy fluctuations carries this mid-level heating farther downstream, resulting in the net cooling seen in Fig. 15e, and it increases the static stability farther downstream (Fig. 13b). This is a separate mechanism by which the stationary circulation is important in shaping the storm tracks. The main contributors in midlatitudes to the reduced vertical temperature gradients and increased static stability are the transient fluxes, with LE flux convergence dominating DSE convergence (cf. Fig. 15a and b). We find that the downstream increase in static stability reduces MAPE by an additional 20% beyond the value of MAPE owing only to reduced meridional temperature gradients by 40% compared with the southern hemisphere. Therefore, modification of temperature gradients through stationary eddies are dominant over modifications of static stability in controlling local baroclinicity in our simulations.

**6. Conclusion**

A common theme that arises in the storm track literature is whether or not the storm track is self-maintained: can the transient eddies maintain themselves in forming the storm track, or are they controlled by other sources such as diabatic heating, surface damping, or stationary waves? Here we have shown that any zonal asymmetry that leads to storm track formation also causes stationary eddies, which both maintain the storm track within half a Rossby wave length of the storm track entrance region, and terminate it farther downstream. Using an aquaplanet moist GCM with localized surface heating, we have shown that the stationary eddies (including interactions with the zonal-mean flow) are locally more important in the energy budget than the transient eddies. The stationary eddy energy flux convergences induced by the zonal asymmetries are large in the storm track region, where they increase temperature gradients, counteracting the tendency of the transient eddies to reduce temperature gradients. Farther downstream, stationary eddy energy fluxes reduce the temperature gradients. This results in regions downstream of the storm track where the EKE is even lower than it would have been without zonal asymmetries and storm tracks, similar to regions of weak eddy activity observed over continents. Therefore, even without continents, the storm track can be well localized, with length scales similar to those observed on Earth.

Varying the rotation rate of the planet allows us to study the generation of storm tracks systematically. We have shown that as the rotation rate is increased, the length...
Fig. 15. Components of the energy flux convergence. (a) Transient eddy DSE flux convergence $-\nabla \cdot (\mathbf{u}'\mathbf{s}')$. (b) Transient eddy LE flux convergence $-L\nabla \cdot (\mathbf{u}'\mathbf{q}')$. (c) Stationary eddy DSE flux convergence $-\nabla \cdot (\mathbf{u}\mathbf{s}^*)$. (d) Stationary eddy LE flux convergence $-L\nabla \cdot (\mathbf{u}\mathbf{q}^*)$. (e) Advection of stationary eddy DSE energy by the zonal-mean wind $-\nabla \cdot (\mathbf{u}\mathbf{s}^*)$. (f) Advection of stationary LE by the mean wind $-L\nabla \cdot (\mathbf{u}\mathbf{q}^*)$. All fields are zonally averaged over a region 60° in longitude downstream of the localized heating.

of the storm track decreases and scales with the length of the stationary eddies, pointing to the role of stationary eddies in terminating the storm track. This occurs both because of enhanced poleward MSE transport downstream of the storm track, which reduces temperature gradients and therefore baroclinicity, and because of increased static stability resulting from LE release at mid-levels, which also reduces baroclinicity. For a climate similar to present-day Earth’s the reduction in temperature gradients is the dominant of these two processes.

The existence of zonal asymmetries leads to a reduction in transient poleward energy fluxes. Stationary energy fluxes not only compensate this reduction but over compensate it, so that the zonal-mean poleward energy transport is increased, resulting in reduced temperature gradients. Local convergences of LE fluxes and DSE fluxes in the storm track region are of similar magnitude, pointing to the importance of moist processes in the dynamics of storm tracks.

Acknowledgments: This research has been supported by the NOAA Climate and Global Change Postdoctoral Fellowship administered by the University Corporation for Atmospheric Research, by a David and Lucile Packard Fellowship, by NSF grant AGS-1019211, and by a Marie Curie career integration grant CIG-304202. We thank Simona Bordoni, Xavier Levine, Tim Merlis, and Adam Sobel for very helpful discussions during the preparation of this manuscript. The simulations were performed on Caltech’s Division of Geological and Planetary Sciences Dell cluster.

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