

How Does Eddy-Latent Heating Coupling Shape the Equilibrated Extratropical Circulation?

Chanil Park^{1,2,3,4}, Paul A. O'Gorman⁵, Seok-Woo Son⁶, Seo-Yeon Kim^{3,4}, Yi Ming^{3,4}

1. Department of Atmospheric Science, University of Alaska Fairbanks

2. International Arctic Research Center, University of Alaska Fairbanks

3. Department of Earth and Environmental Sciences, Boston College

4. Boston College

5. Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology

6. School of Earth and Environmental Sciences, Seoul National University

Abstract

Latent heating (LH) influences the extratropical large-scale circulation, but its dynamical role remains difficult to isolate because moist processes simultaneously affect eddies, the zonal-mean state, and their interaction. Here, we investigate how eddy-LH coupling shapes the extratropical circulation in statistical equilibrium by conducting a series of moist idealized general circulation model experiments in which the coupling between LH and eddies is selectively suppressed while preserving the direct effect of LH on the zonal-mean state. As eddy-LH coupling is progressively suppressed, the effective static stability experienced by eddies approaches the dry static stability. The zonal-mean state concurrently undergoes a systematic reorganization characterized by an enhanced meridional temperature gradient in midlatitudes associated with high-latitude cooling and a poleward intensification of the eddy-driven jet. The meridional temperature gradient increases quasi-linearly with the effective static stability, reminiscent of baroclinic adjustment in a moist atmosphere. Despite the loss of diabatic eddy energy input, storm track activity does not weaken owing to the compensating increase in baroclinic energy conversion associated with the enhanced meridional temperature gradient. The dominant length scale of extratropical eddies increases systematically as eddy-LH coupling weakens, consistent with scaling based on the Rossby deformation radius defined using the effective static stability. These results demonstrate that the role of eddy-LH coupling in the equilibrated extratropical circulation extends beyond its established influence on linear moist baroclinic instability to include the regulation of the background mean state.

How Does Eddy-Latent Heating Coupling Shape the Equilibrated Extratropical Circulation?

Chanil Park^{1,2,3,4*}, Paul A. O'Gorman⁵, Seok-Woo Son⁶, Seo-Yeon Kim^{3,4}, and Yi Ming^{3,4}

¹ *Department of Atmospheric Science, University of Alaska Fairbanks, Fairbanks, Alaska, United States*

² *International Arctic Research Center, University of Alaska Fairbanks, Fairbanks, Alaska, United States*

³ *Department of Earth and Environmental Sciences, Boston College, Chestnut Hill, Massachusetts, United States*

⁴ *Schiller Institute for Integrated Science and Society, Boston College, Chestnut Hill, Massachusetts, United States*

⁵ *Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts, United States*

⁶ *School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea*

* **Corresponding author:** Chanil Park (Email: cpark15@alaska.edu)

Abstract

Latent heating (LH) influences the extratropical large-scale circulation, but its dynamical role remains difficult to isolate because moist processes simultaneously affect eddies, the zonal-mean state, and their interaction. Here, we investigate how eddy-LH coupling shapes the extratropical circulation in statistical equilibrium by conducting a series of moist idealized general circulation model experiments in which the coupling between LH and eddies is selectively suppressed while preserving the direct effect of LH on the zonal-mean state. As eddy-LH coupling is progressively suppressed, the effective static stability experienced by eddies approaches the dry static stability. The zonal-mean state concurrently undergoes a systematic reorganization characterized by an enhanced meridional temperature gradient in midlatitudes associated with high-latitude cooling and a poleward intensification of the eddy-driven jet. The meridional temperature gradient increases quasi-linearly with the effective static stability, reminiscent of baroclinic adjustment in a moist atmosphere. Despite the loss of diabatic eddy energy input, storm track activity does not weaken owing to the compensating increase in baroclinic energy conversion associated with the enhanced meridional temperature gradient. The dominant length scale of extratropical eddies increases systematically as eddy-LH coupling weakens, consistent with scaling based on the Rossby deformation radius defined using the effective static stability. These results demonstrate that the role of eddy-LH coupling in the equilibrated extratropical circulation extends beyond its established influence on linear moist baroclinic instability to include the regulation of the background mean state.

1. Introduction

The extratropical large-scale circulation is fundamentally driven by the equator-to-pole temperature contrast, which provides the reservoir of mean available potential energy (MAPE) that can be converted into eddy available potential energy (EAPE) and then into eddy kinetic energy (EKE) (Lorenz 1955; Peixóto and Oort 1974). This conversion is achieved through baroclinic instability (Charney 1947; Eady 1949; Phillips 1954), whose large-scale effects are reflected in eddy heat fluxes (Lorenz 1955; Stone 1978; Held and Larichev 1996), and manifest themselves on synoptic weather maps as surface cyclogenesis and its vertical coupling with upper-tropospheric disturbances (Petterssen and Smebye 1971; Hoskins et al. 1985). Baroclinically growing eddies exhibit a distinct lifecycle (Thorncroft et al. 1993) and interact with the mean flow, driving the eddy-driven jet (EDJ) and Ferrel circulation (Edmon et al. 1980; Robinson 2000) and shaping the thermal stratification (Stone 1978; Schneider 2004). These dry dynamical processes form the geophysical fluid dynamical backbone of our understanding of the extratropical climate.

Although long regarded as secondary to dry dynamics, moist processes also play an important role in shaping the extratropical weather and climate. However, unlike the tropics, where moist convection and the large-scale circulation are inherently tightly coupled (Riehl and Malkus 1958; Emanuel et al. 1994), moist dynamics has received relatively less attention in the extratropics. Given the increasing moisture content and its expected effects in weather and climate under anthropogenic warming (O’Gorman and Schneider 2009; Schneider et al. 2010), improving our understanding of extratropical moist dynamics becomes an increasingly important challenge.

The existing body of literature on extratropical moist dynamics is broadly twofold: studies examining the effects of latent heating (LH) on synoptic storms and those examining its effects on the large-scale mean state (pathways A and B, respectively, in Fig. 1). The effects of LH on synoptic storms have been investigated through the development of moist baroclinic instability theories (Wang and Barcilon 1986; Mak 1982; Emanuel et al. 1987; Mak 1994; Kohl and O’Gorman 2022; Yang and Li 2023), as well as through diagnostic approaches in which LH effects can be explicitly quantified for individual storms (Davis and Emanuel 1991; Stoelinga 1996; Tamarin and Kaspi 2016; Zhao et al. 2023; Park et al. 2023). Linear theories predict that LH increases the growth rate of storms and preferentially destabilizes shorter waves. This is well supported by moist baroclinic wave lifecycle experiments (Gutowski et al. 1992; Whitaker and Davis 1994; Balasubramanian and Yau 1996; Moore and Montgomery 2004;

Booth et al. 2013, 2015), although LH may not lead to further storm amplification at higher background temperatures due to degraded phasing between eddy fields (Kirshbaum et al. 2018). For a comprehensive review, readers are referred to Wernli and Gray (2024).

The aforementioned studies primarily address the linear growth or finite-time evolution of baroclinic disturbances, for which the background mean state is either prescribed or undergoes little adjustment. Moist processes, however, also directly modify the mean state in which baroclinic disturbances develop. As in the tropics (Xu and Emanuel 1989), moist convection contributes to shaping the static stability and tropopause in the extratropics (Juckes 2000; Frierson 2008; Schneider and O’Gorman 2008; Wu and Pauluis 2014; Levine and Schneider 2015). Studies have also shown that LH exerts dynamical influences on the extratropical large-scale circulation, including the EDJ and Ferrel cell (Lachmy and Kaspi 2020; Lachmy 2022; Ghosh et al. 2024), and their low-frequency variability (Xia and Chang 2014; Lutsko and Hell 2021). In these studies, however, the role of LH is interpreted mostly through its statistical imprint left by weather systems (e.g., zonal-mean LH), rather than through its explicit coupling with eddies. Some studies have employed dry dynamical-core general circulation models (GCMs) with prescribed zonally uniform thermal forcings to investigate the dynamical role of diabatic heating in the extratropical large-scale circulation (Baker et al. 2017; White et al. 2024).

Taken together, LH influences the extratropical circulation by modifying both eddies and the mean state. In addition to these direct influences, LH may further indirectly affect dry eddy-mean flow interaction (pathway C in Fig. 1). For example, LH-induced changes in storm activity can alter meridional eddy heat transport and thereby modify the equator-to-pole temperature gradient. Conversely, changes in the mean thermal structure can feed back onto eddies; for example, the greater dry static stability of a moist atmosphere can suppress EKE (Lutsko et al. 2024). However, the consequences of these individual effects for the equilibrated circulation are difficult to predict because of nonlinear interactions between all the players in Fig. 1. A few studies have considered all these influence pathways in interpreting the circulation responses using numerical experiments in which atmospheric moisture is varied and the mean state is allowed to adjust freely (Frierson et al. 2006; Bembenek et al. 2020; Lutsko et al. 2024). For example, Lutsko et al. (2024) showed that increasing atmospheric moisture can ultimately weaken EKE when the environmental conditions are permitted to adjust. However, the individual influence pathways have rarely been isolated thus far, limiting confidence in our systematic understanding of moist dynamics. In particular, the role of eddy-

LH coupling in shaping the extratropical circulation beyond its well-established role in moist baroclinic instability remains unclear.

In the present study, we explore how eddy-LH coupling shapes the statistically equilibrated state of the extratropical circulation using an idealized moist GCM coupled to a slab ocean. To isolate the role of eddy-LH coupling from the multiple influence pathways through which moisture influences the extratropical circulation, we introduce a LH homogenization framework that selectively suppresses the zonally asymmetric component of LH while preserving the direct effect of LH on the zonal-mean state. Within this framework, the effective static stability experienced by eddies approaches the dry static stability, providing a simple physical interpretation in terms of dry dynamical arguments. The strength of eddy-LH coupling is then varied systematically using a tunable homogenization parameter to trace the circulation response across coupling strengths. Specifically, we aim to answer the following questions using this framework:

- 1) How does the zonal-mean background state adjust when eddy-LH coupling is suppressed?
- 2) Does storm track activity weaken, as might be expected from moist baroclinic instability theory, when eddy-LH coupling is suppressed but mean-field adjustment is allowed?
- 3) Does the dominant length scale of extratropical eddies increase when eddy-LH coupling is suppressed, as expected from the comparison of the most unstable modes of dry versus moist baroclinic instability?

The remainder of this paper is organized as follows. Section 2 describes the idealized moist GCM and the experimental design, including the LH homogenization framework. Section 3 presents the theoretical basis of the LH homogenization in terms of the effective static stability. The three questions posed above are addressed in section 4. Section 5 summarizes the main findings and discusses their broader implications.

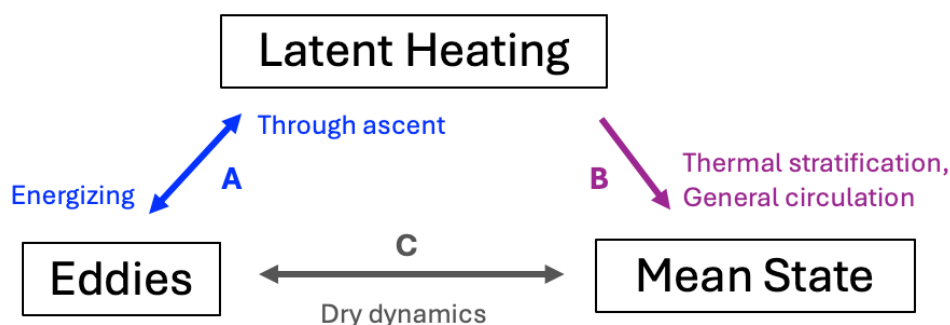


Figure 1. Conceptual illustration of the influence pathways linking latent heating, eddies, and the mean state in the extratropical atmosphere. Single-headed arrows denote one-way influences, whereas the double-headed arrow denotes a two-way interaction.

2. Methods

a. Model description

We use an idealized moist GCM configured similarly to the Model of an Idealized Moist Atmosphere (MiMA; Jucker and Gerber 2017), which is built upon the aquaplanet GCM of Frierson et al. (2006) and implemented within the Isca modeling framework (Vallis et al. 2018). It employs a spectral dynamic core solving the primitive equations in sigma coordinates. The surface is represented by a slab ocean whose temperature evolves according to local energy exchanges with the atmosphere as

$$C \frac{\partial T_{slab}}{\partial t} = R_{net} + H_L + H_S, \quad (1)$$

where C is the heat capacity of the slab ocean, a function of the mixed-layer depth; R_{net} is the net surface radiative flux; H_L and H_S are the surface latent and sensible heat fluxes, respectively.

Moist convection is parameterized using a simplified Betts-Miller convection scheme (Betts and Miller 1986; Frierson 2007) with minor modifications of O’Gorman and Schneider (2008). A large-scale condensation scheme produces resolved precipitation when the air becomes saturated. The condensed water is assumed to precipitate immediately, with no explicit representation of clouds. Only vapor-liquid phase changes are considered, and re-evaporation is not permitted.

Radiative heating is computed using the Rapid Radiative Transfer Model (RRTM; Mlawer et al. 1997). H₂O, CO₂, and O₃ are the only radiative active gases represented. H₂O is computed interactively through the model hydrological cycle, whereas CO₂ and O₃ are prescribed. Surface fluxes are calculated using bulk aerodynamic formulas. Their drag coefficients are derived from the modified Monin–Obukhov scaling used in O’Gorman and Schneider (2008) which allows for unstable boundary layers. Turbulent mixing within the planetary boundary layer is parameterized using a K-profile diffusion scheme with a boundary layer depth diagnosed from a critical bulk Richardson number (Troen and Mahrt 1986).

b. Core experimental designs

We run the model at T42 horizontal resolution with 40 unevenly spaced sigma levels. A higher-resolution configuration (T85) was also tested and yielded robust conclusions; therefore, T42 results are presented here to facilitate reproduction at lower computational cost. The slab ocean has a depth of 1 m and a globally uniform surface albedo of 0.26. The moist convective adjustment time scale and reference relative humidity are set to 2 h and 0.8, respectively. CO₂ is prescribed as a spatially uniform concentration of 370 ppm, while ozone is specified using a climatological vertical-latitude distribution. The radiative time step is 2 h.

In the absence of ocean heat transport, all meridional heat transport is accomplished by the atmosphere. This produces an excessively strong meridional temperature gradient in the tropics, leading to an unrealistically strong Hadley circulation and subtropical jet. An overly strong subtropical jet can further result in too equatorward positions of EDJ and storm track (Lee and Kim 2003; Son and Lee 2005; Brayshaw et al. 2008). To alleviate this, we introduce an additional term, $\nabla \cdot \mathbf{F}_o$, into the right-hand side of Eq. (1) to mimic the ocean heat transport (the so-called Q flux), following Merlis et al. (2013):

$$\nabla \cdot \mathbf{F}_o(\phi) = Q_o \frac{1}{\cos\phi} \left(1 - \frac{2\phi^2}{\phi_0^2}\right) \exp\left(-\frac{\phi^2}{\phi_0^2}\right), \quad (2)$$

where ϕ is the latitude; Q_o and ϕ_0 are set to 35 W m⁻² and 16°, respectively. This Q flux effectively redistributes excess heat from the deep tropics to the subtropics (Jucker and Gerber 2017; Garfinkel et al. 2020).

The model is integrated under a perpetual equinox condition with no seasonal or diurnal cycle. All simulations are conducted for 7,200 days, with the first 3,600 days discarded as spinup. Statistical equilibrium states are computed from the final 3,600 days. Given the hemispheric symmetry of the equilibrated climate, the Northern and Southern Hemispheres are combined. Model output, originally defined in native sigma coordinates, is interpolated onto standard pressure levels before analysis.

c. Latent-heating homogenization experiments

To investigate the influence of eddy-LH coupling on the equilibrated extratropical circulation, we conduct LH homogenization experiments. At every model time step, the diabatic temperature tendency due to LH, $\dot{T}_{LH}$, is zonally homogenized according to

$$\dot{T}_{LH} = \overline{\dot{T}_{LH}} + \alpha \dot{T}'_{LH}, \quad (3)$$

where overbar and prime denote zonal mean and deviations therefrom, respectively. Here, α is a tunable parameter controlling the extent of homogenization. When $\alpha = 1.0$, LH at each grid point is identical to that computed by the moist convection and large-scale condensation schemes. When $\alpha = 0.0$, LH is replaced by its zonal mean. Those two experiments are referred to as control and homogenization experiments, respectively.

We perform parameter-sweep experiments in which α is varied from 0.0 to 1.0 in increments of 0.1 to examine how the equilibrated extratropical circulation responds to varying degrees of eddy-LH coupling suppression. Note that regardless of the value of α , the zonal-mean LH is preserved, allowing the effects of eddy-LH coupling to be isolated without altering the direct effect of LH in zonal-mean dynamics. To focus on the role of eddy-LH coupling in the extratropics, LH is homogenized only poleward of 30° . Artificial discontinuities between the tropics and extratropics are avoided by allowing the homogenization parameter to vary smoothly across the subtropics using a cosine taper:

$$\alpha_\phi = \begin{cases} 1, & |\phi| < \phi_1 \\ \alpha + \frac{1-\alpha}{2} \left[1 + \cos\left(\pi \frac{|\phi| - \phi_1}{\phi_2 - \phi_1}\right) \right], & \phi_1 \leq |\phi| \leq \phi_2, \\ \alpha, & |\phi| > \phi_2 \end{cases} \quad (4)$$

with $\phi_1 = 20^\circ$ and $\phi_2 = 30^\circ$.

Figure 2 shows snapshots of $\dot{T}_{LH}$ at 500 hPa and geopotential height anomaly from its zonal mean at 750 hPa for the control ($\alpha = 1.0$) and homogenization ($\alpha = 0.0$) experiments. In the control experiment, LH occurs preferentially along the leading edge of cyclonic anomalies (Fig. 2a), consistent with the observed midlatitude synoptic storms. In the homogenization experiment, $\dot{T}_{LH}$ in the extratropics exhibits no zonal asymmetry (Fig. 2b), thereby eliminating spatial correlations between LH and eddy circulation. The next section further details the theoretical basis of the LH homogenization framework.

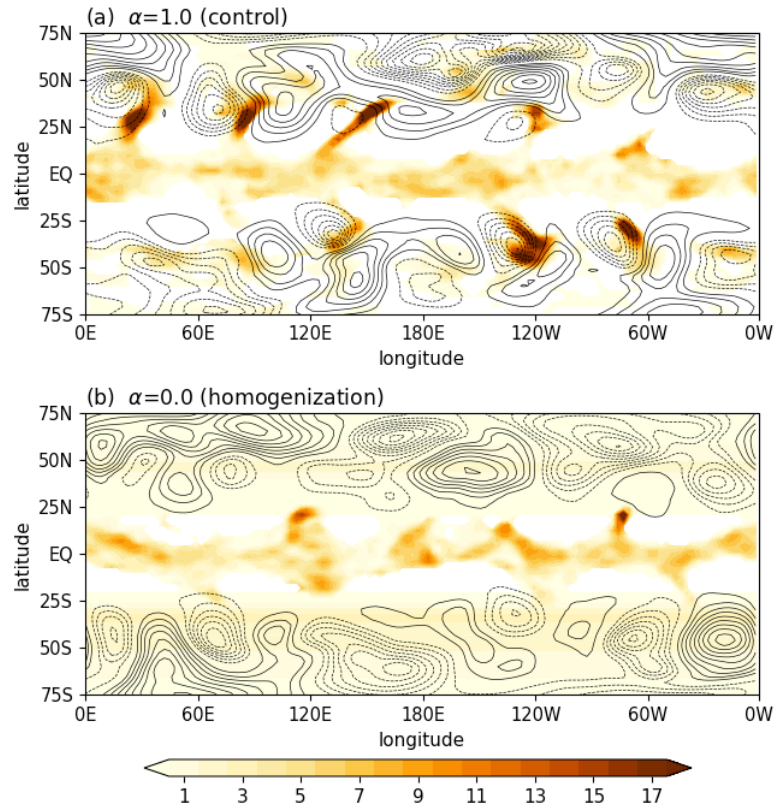


Figure 2. Snapshots of instantaneous zonal anomaly of 700-hPa geopotential height (contours, unit: gpm) and 500-hPa temperature tendency due to LH (shading, unit: K day⁻¹) in **a** the control ($\alpha=1.0$) and **b** the homogenization ($\alpha=0.0$) experiments.

3. Theoretical basis of LH homogenization

One intuitive way to understand the impact of LH homogenization may be through potential vorticity (PV) thinking (Hoskins et al. 1985). In midlatitude storms, latent heat release is collocated with ascending motion on the downstream side of low-level cyclones (as shown in Fig. 2a), generating upper-level anticyclonic and low-level cyclonic PV anomalies through altered dry static stability ($\partial\theta/\partial p$, where θ is potential temperature). These PV anomalies are balanced by upper-level ridge building and low-level cyclone intensification, thereby reinforcing the mutual interaction between upper- and low-level baroclinic disturbances when the phasing of LH is favorable. Homogenizing LH destroys this phase relationship, preventing LH from amplifying baroclinic disturbances.

Although PV thinking provides an intuitive picture of the consequence of LH homogenization for individual storms, it offers limited insight into how these effects accumulate to shape the equilibrated eddies and mean fields. To interpret the gross influence

of eddy-LH decoupling in extratropical circulation, we adopt the effective static stability theory of O’Gorman (2011). In the homogenization experiment ($\alpha = 0.0$), the effective static stability felt by eddies reduces to dry static stability, implying that eddies follow the dry dynamics even in the presence of moist processes. To understand how this emerges, we first revisit the derivation of the effective static stability.

For an air parcel undergoing condensation while conserving its saturation equivalent potential temperature, θ_e^* , the thermodynamic equation can be written as

$$\frac{D\theta}{Dt} = \omega^\uparrow \left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*} = \dot{T}_{LH}, \quad (5)$$

where $\omega^\uparrow = \omega H(-\omega)$ and $H(\cdot)$ is the Heaviside function, which ensures that condensation occurs only during ascent. It is assumed that LH is the dominant diabatic process on storm time scale, and that the ascending air parcel remains saturated. Transforming Eq. (5) into Eulerian form and collecting the terms involving vertical velocity, ω , yields

$$\frac{\partial \theta}{\partial t} = -\omega \frac{\partial \theta}{\partial p} + \omega^\uparrow \left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*} + \dots, \quad (6)$$

Neglecting the eddy contribution to the vertical gradient of potential temperature (i.e., $\partial \theta / \partial p \approx \partial \bar{\theta} / \partial p$), the eddy thermodynamic equation is

$$\frac{\partial \theta'}{\partial t} = -\omega' \frac{\partial \bar{\theta}}{\partial p} + \omega'^\uparrow \overline{\left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*}} + \dots. \quad (7)$$

The truncated eddy ascent $\omega'^\uparrow$ is then parameterized in terms of ω' using a least-squares regression in the zonal direction:

$$\omega'^\uparrow = \lambda \omega' + \varepsilon, \quad (8)$$

where $\lambda = \overline{\omega' \omega'^\uparrow} / \overline{\omega'^2}$ is the rescaling coefficient; ε is the residual. To the extent that $\omega'^\uparrow$ is reasonably projected onto ω' , Eq. (7) is rewritten as

$$\frac{\partial \theta'}{\partial t} = -\omega' \left(\frac{\partial \bar{\theta}}{\partial p} \right)_{\text{eff}} + \dots, \quad (9)$$

where the effective static stability is defined by

$$-\left(\frac{\partial \bar{\theta}}{\partial p} \right)_{\text{eff}} = -\frac{\partial \bar{\theta}}{\partial p} + \lambda \overline{\left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*}}. \quad (10)$$

Note that Eq. (9) is identical to the dry eddy thermodynamic equation except that the dry static stability is replaced by the effective static stability. This formulation, although post hoc, allows

dry eddy dynamics to be applied to a moist atmosphere (O’Gorman 2011; Levine and Schneider 2015; Dwyer and O’Gorman 2017), which would not be possible with previous reduced static stability that does not take unsaturated descent into account (Nielsen-Gammon and Keyser 2000; Zurita-Gotor 2005; Emanuel et al. 1987). We now extend the above derivation by replacing the LH term with its homogenized form. Equation (6) is then modified as

$$\frac{\partial \theta}{\partial t} = -\omega \frac{\partial \theta}{\partial p} + \overline{\omega^\uparrow \left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*}} + \alpha \left[\omega^\uparrow \left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*} - \overline{\omega^\uparrow \left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*}} \right] + \dots, \quad (11)$$

where the second and third terms on the right-hand side correspond to the right-hand side of Eq. (3). Likewise, the eddy thermodynamic equation is

$$\frac{\partial \theta'}{\partial t} = -\omega' \frac{\partial \bar{\theta}}{\partial p} + \alpha \omega'^\uparrow \overline{\left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*}} + \dots, \quad (12)$$

which can be rewritten in a similar way as

$$\frac{\partial \theta'}{\partial t} = -\omega' \left(\frac{\partial \bar{\theta}}{\partial p} \right)_{eff} + \dots, \quad (13)$$

where the effective static stability is now

$$-\left(\frac{\partial \bar{\theta}}{\partial p} \right)_{eff} = -\frac{\partial \bar{\theta}}{\partial p} + \alpha \lambda \overline{\left(\frac{\partial \theta}{\partial p} \right)_{\theta_e^*}}. \quad (14)$$

Equation (14) states that LH homogenization modifies the effective static stability in a simple manner: the stratification effect by LH is merely rescaled by α . Consequently, the effective static stability reduces to the dry static stability in the limit of complete homogenization ($\alpha = 0$), whereas it recovers the original expression when $\alpha = 1$.

The rescaling coefficient λ is generally less than unity because Eq. (8) accounts for both condensation during ascent and its absence during descent. At latitudes of strong zonal-mean ascent (e.g., deep tropics), λ is close to 1, whereas λ is close to 0 at latitudes of strong zonal-mean descent (e.g., subtropics). At latitudes where zonal-mean vertical velocity is close to zero (e.g., midlatitudes), λ measures the asymmetry of ω' with $\lambda = 0.5$ for symmetry between ascent and descent, and $\lambda > 0.5$ for ascent stronger and spatially narrower than descent. See O’Gorman (2011) for further interpretation.

In the control experiment, λ is close to 0.8 in the deep tropics, while it is less than 0.5 in the subtropics (Fig. 3a). In midlatitudes, λ exceeds 0.5, implying that the distribution of

vertical velocity is skewed toward stronger and narrower ascent. This skewness is a common feature of moist baroclinic disturbances, in which latent heat release and ascent mutually reinforce each other, leading to stronger and more localized upward motion on the frontal scale (Emanuel et al. 1987; Booth et al. 2015). The homogenization experiment exhibits a pronounced reduction in λ in midlatitudes (Fig. 3b). This reduction reflects the removal of the positive feedback between LH and ascent, which are no longer spatially correlated to each other. In contrast, λ changes little in the deep tropics where homogenization is not applied and in the high latitudes where the amount of moisture is low. The reduction is most prominent in the lower troposphere in storm track latitudes and rapidly weakens above the tropopause. The tropopause height does not change much in the homogenization experiment (not shown). When averaged over the midlatitudes and whole troposphere, λ linearly decreases as α decreases from 1.0 to 0.0 (Fig. 4a), reflecting a gradual weakening of the ascent-descent asymmetry within baroclinic eddies.

The LH homogenization substantially increases the effective static stability throughout the extratropical troposphere (Figs. 3c,d). The largest increases occur in the lower troposphere within the storm track latitudes, roughly coinciding with the region where the asymmetry parameter also exhibits its largest reduction. This spatial correspondence suggests that this increase in effective static stability is primarily driven by suppressed coupling between eddies and LH, rather than by changes in the dry static stability. Indeed, the dry static stability exhibits modest changes of varying sign (Figs. 3e,f). Near the tropopause and in the lower stratosphere, however, the responses of effective and dry static stabilities are very similar, suggesting that the effective static stability change there is dominated by adjustments of the dry stratification. Averaged over the midlatitudes and troposphere, the effective static stability increases nearly linearly as α decreases, approaching the dry static stability at $\alpha=0.0$ (Fig. 4b). In contrast, the dry static stability remains nearly unchanged across the parameter sweep of α . If atmospheric moisture content, rather than eddy-LH coupling, were altered, the dry static stability would change substantially as shown by Frierson et al. (2006) and Lutsko et al. (2024). Therefore, the response of the effective static stability is governed primarily by changes in eddy-LH coupling rather than by changes in the background thermal stratification.

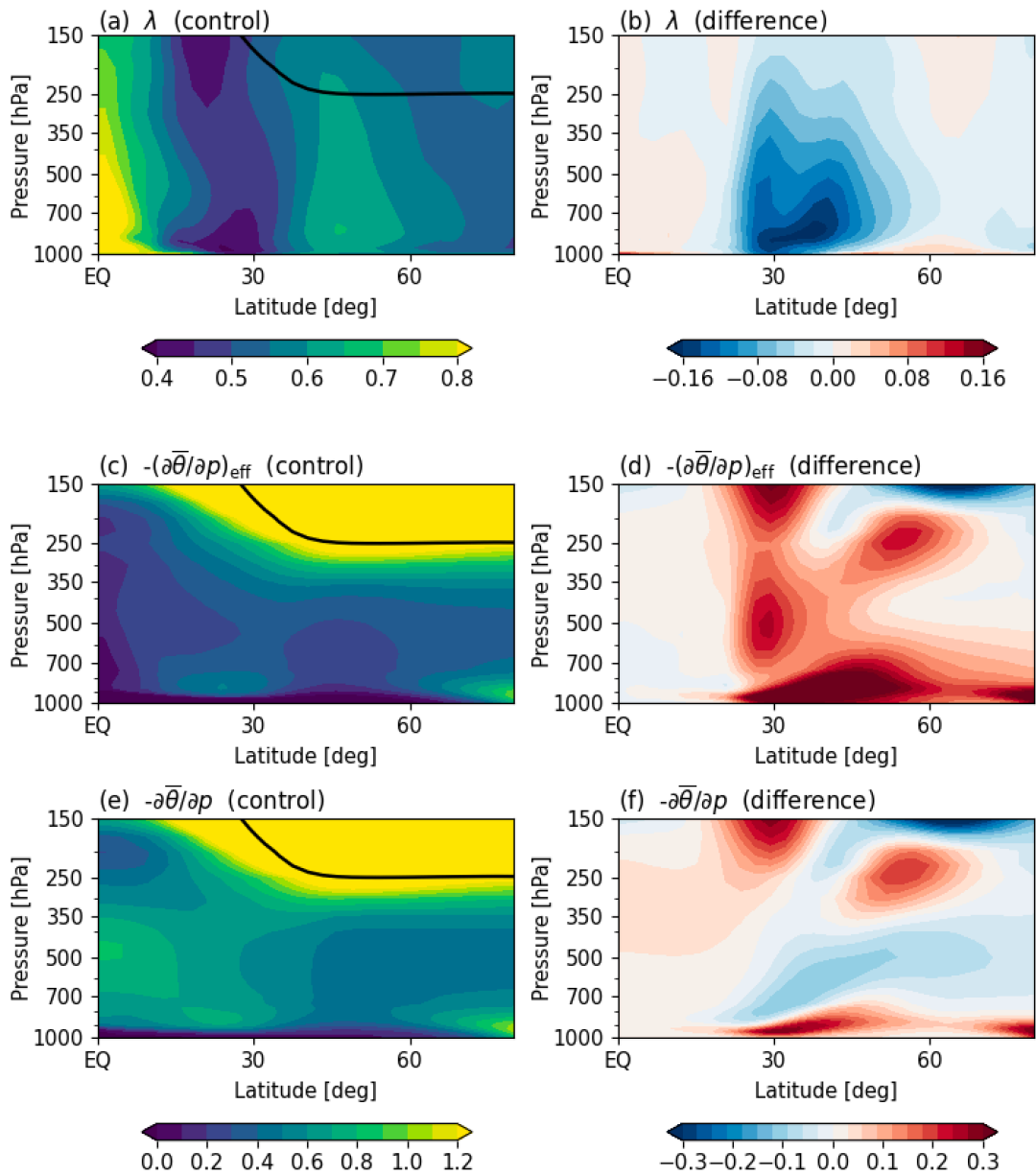


Figure 3. **a** λ (shading) and dynamic tropopause (black lines) in the control experiment. The dynamic tropopause is defined as the 2-PVU surface, where $1 \text{ PVU} \equiv 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$. **b** Difference in λ between the homogenization and control experiments ($\alpha=0.0$ minus $\alpha=1.0$). **c,d**, as in **a,b**, but for effective static stability (unit: K hPa^{-1}). **e,f**, as in **a,b**, but for dry static stability (unit: K hPa^{-1}).

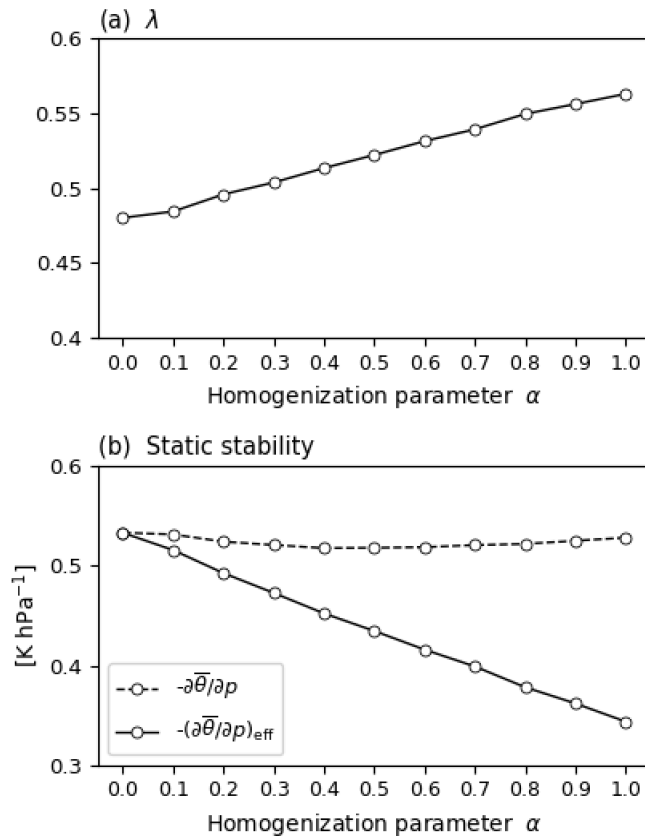


Figure 4. **a** Tropospheric (1000–250 hPa) and midlatitude (30°–60°) average of λ for a series of experiments in which α is varied from 0.0 to 1.0. **b** As in **a**, but for dry static stability (dashed line) and effective static stability (solid line).

4. Equilibrated extratropical large-scale circulation

a. Adjustment of mean fields

We examine first how the zonal-mean fields adjust to changes in eddy-LH coupling. Relative to the control, the homogenization experiment exhibits a colder atmosphere poleward of 50° and a warmer lower troposphere on the equatorward side of the storm track, thereby enhancing the meridional thermal contrast in midlatitudes (Fig. 5a). Consistent with thermal wind balance, the strengthened temperature gradient is accompanied by a poleward displacement and intensification of the westerly jet (Fig. 5b). The jet response is nearly barotropic, implying the change in the eddy-driven jet, rather than in the angular momentum-conserving subtropical jet that is located on the poleward edge of the Hadley cell (Held and Hou 1980). Indeed, we found that the latitude of the Hadley cell edge remains nearly unchanged, while the eddy momentum flux convergence varies consistently with the jet change (not shown).

The enhanced meridional thermal gradient is also reflected in the slab-ocean temperature, which evolves through local atmosphere-ocean energy exchange. As the homogenization parameter α decreases from 1.0 to 0.0, high-latitude surface temperature decreases systematically, whereas tropical surface temperature remains nearly unchanged (Fig. 5c). The corresponding response of the EDJ position, defined as the latitude of the maximum zonal-mean zonal wind at 850 hPa, is provided in Fig. 5d. The jet shifts progressively poleward with decreasing α , with the largest sensitivity occurring for $\alpha \leq 0.4$. When $\alpha = 0.0$, the EDJ is displaced poleward by approximately 4° relative to the control. The more causal mechanisms of EDJ response to eddy-LH decoupling are beyond the scope of the present study.

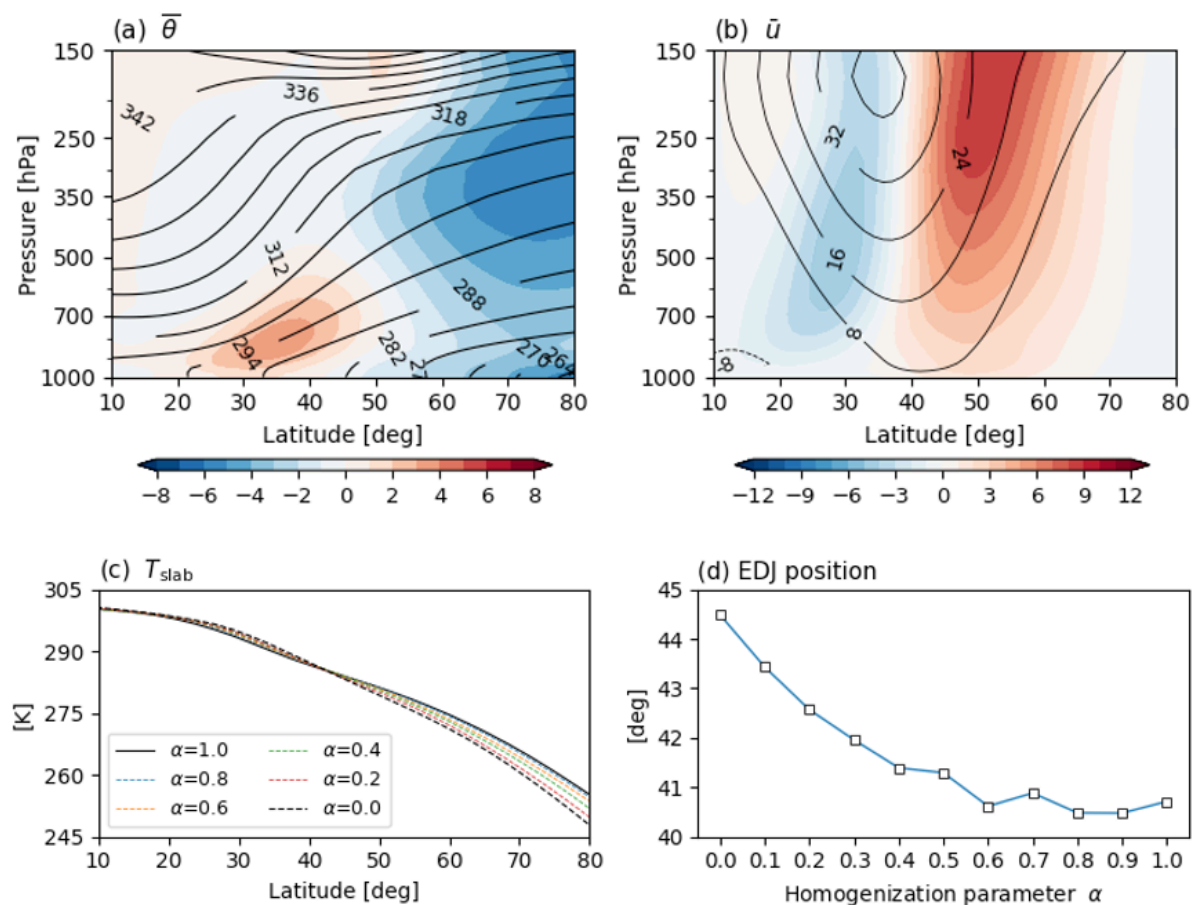


Figure 5. **a** Zonal-mean potential temperature in the control experiment (contours, unit: K) and its difference between the homogenization and control experiments ($\alpha=0.0$ minus $\alpha=1.0$; shading). **b** As in **a**, but for zonal-mean zonal wind (unit: m s^{-1}). **c** Latitudinal profiles of zonal-mean slab ocean surface temperature, and **d** latitudinal positions of EDJ for a series of experiments in which α is varied from 0.0 to 1.0.

Because the zonal-mean thermal structure shown in Fig. 5 represents an equilibrated state arising from nonlinear interactions between eddies and the mean flow over model integration, direct attribution of the mean-state adjustment to eddy-LH coupling is not

straightforward. Instead, we examine whether the adjusted thermal structure can be understood using scaling arguments. We first consider the baroclinic adjustment theory of Stone (1978). The original baroclinic adjustment states that baroclinic eddies continuously adjust the extratropical troposphere toward a state that is neutral to linear baroclinic instability. The resulting scaling relates the dry static stability to the meridional temperature gradient as $\frac{\partial \bar{\theta}}{\partial z} = -\frac{f}{\beta H} \frac{\partial \bar{\theta}}{\partial y}$, where f and β are the planetary vorticity and its meridional gradient, and H is the scale height. This scaling suggests that a baroclinically adjusted atmosphere maintains an isentropic slope at its critical value. However, statistical equilibrium in nonlinear models does not necessarily coincide with exact baroclinic neutrality (Zhou and Stone 1993a,b; Barry et al. 2000). Nevertheless, a tendency for eddies to constrain variations in the isentropic slope is evident in the real atmosphere, as indicated by the lead-lag relationship between anomalous isentropic slope and synoptic eddy heat fluxes (Thompson and Birner 2012). Following the concept of “weak baroclinic adjustment” (Stone and Branscome 1992), the scaling relevant to our parameter-sweep experiments may be expressed as

$$\frac{\partial \bar{\theta}}{\partial z} \propto -\frac{\partial \bar{\theta}}{\partial y}. \quad (15)$$

An analogous scaling for a moist atmosphere may be obtained by replacing the dry static stability with the effective static stability:

$$\left(\frac{\partial \bar{\theta}}{\partial z}\right)_{\text{eff}} \propto -\frac{\partial \bar{\theta}}{\partial y}. \quad (16)$$

Figure 6 compares the weak baroclinic adjustment scaling with its moist extension in our parameter-sweep experiments. Both sides of each equation are evaluated at 500 hPa and averaged over the midlatitudes (30°–60°). The hydrostatic approximation (i.e., $\partial p = -\rho g \partial z$) is used to express the static stabilities in geometric coordinates. Across the parameter sweep, the effective static stability varies quasi-linearly with the meridional potential temperature gradient. Thus, the systematic increase in the effective static stability with decreasing α is accompanied by a corresponding strengthening of the meridional temperature gradient, broadly consistent with baroclinic adjustment behavior. In contrast, the dry static stability varies weakly and in the opposite direction. This decrease differs from the nearly constant dry static stability in Fig. 4b because the former is evaluated at a single level rather than averaged over the troposphere.

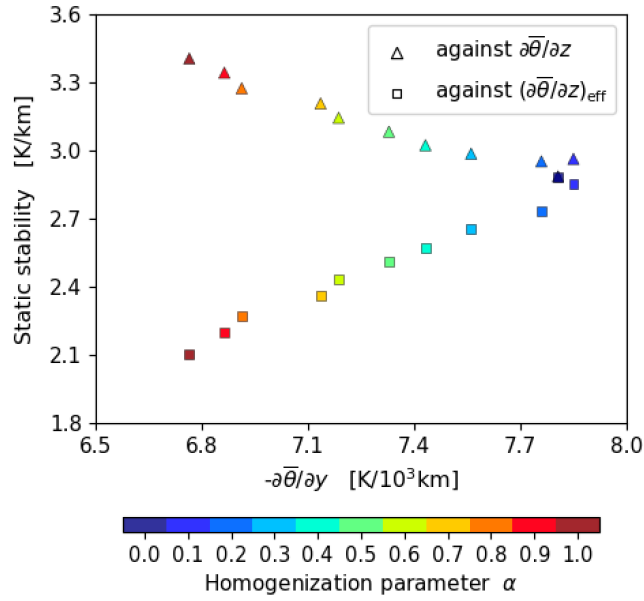


Figure 6. Meridional potential temperature gradient as a function of midlatitude (30° – 60°) dry static stability (triangles) and effective static stability (squares) at 500 hPa for a series of experiments in which α is varied from 0.0 to 1.0.

While the weak baroclinic adjustment scaling qualitatively accounts for the covariation between effective static stability and the meridional temperature gradient, the unspecified proportionality coefficient precludes a quantitative prediction of their magnitudes. We next examine whether the supercriticality framework of Schneider and Walker (2006) provides a quantitative constraint on the adjusted thermal structure. The supercriticality is expressed as

$$S = \frac{f}{\beta} \frac{\partial \bar{\theta}_S}{\partial y} \left[2 \frac{\partial \bar{\theta}_S}{\partial p} (\bar{p}_S - \bar{p}_T) \right]^{-1}, \quad (17)$$

where $\frac{\partial \bar{\theta}_S}{\partial y}$ and $\frac{\partial \bar{\theta}_S}{\partial p}$ are the near-surface meridional potential temperature gradient and dry static stability; $\bar{p}_S - \bar{p}_T$ is the pressure depth of the troposphere. The supercriticality S quantifies the ratio of the pressure depth over which baroclinic eddies redistribute entropy to the total pressure depth of the troposphere. If baroclinic eddies dominate the tropospheric entropy redistribution and thermal stratification, S is expected to be close to 1. When $S < 1$, baroclinic eddies do not extend throughout the full depth of the troposphere, and the upper-tropospheric thermal stratification is primarily maintained by radiation and convection. The latter regime does not generally occur unless the near-surface baroclinicity is very weak (Schneider and Walker 2006; Levine and Schneider 2015). Although formally similar to the original baroclinic adjustment condition, Eq. (17) does not assume baroclinic neutrality but instead follows from constraints on isentropic mass transport in a continuously stratified atmosphere (Schneider 2004).

O’Gorman (2011) extended this framework to moist atmospheres by replacing the dry static stability with the effective static stability:

$$S_{\text{eff}} = \frac{f}{\beta} \frac{\partial \bar{\theta}_S}{\partial y} \left[2 \left(\frac{\partial \bar{\theta}_S}{\partial p} \right)_{\text{eff}} (\bar{p}_S - \bar{p}_T) \right]^{-1}, \quad (18)$$

that can be used to constrain the moist extratropical atmosphere. This extension is advocated by Levine and Schneider (2015) who demonstrated that S_{eff} remains close to 1 across a range of moist idealized GCM configurations, just as S does in dry atmospheres.

Figure 7 shows the supercriticality S and its moist extension S_{eff} in our parameter-sweep experiments. $\frac{\partial \bar{\theta}_S}{\partial y}$ and static stabilities are evaluated at 875 hPa, immediately above the boundary layer top, and averaged over midlatitudes (30°–60°). The pressure depth $\bar{p}_S - \bar{p}_T$ is determined as the difference between 875 hPa and the mean pressure of the 2-PVU surface. S has values around 0.5 across the parameter sweep of α , underestimating the actual vertical extent of baroclinic eddies (Schneider and O’Gorman 2008; Levine and Schneider 2015). S_{eff} has a value smaller than but very close to 1 for the control experiment, consistent with Levine and Schneider (2015). However, it decreases continuously to ~0.5 as α varies from 1.0 to 0.0, rather than remaining close to unity. We also took the average over baroclinic zones defined as a finite latitude width with respect to maximum eddy heat flux, instead of fixed 30°–60°, as in previous studies (O’Gorman 2011; Levine and Schneider 2015), but the results remain unchanged.

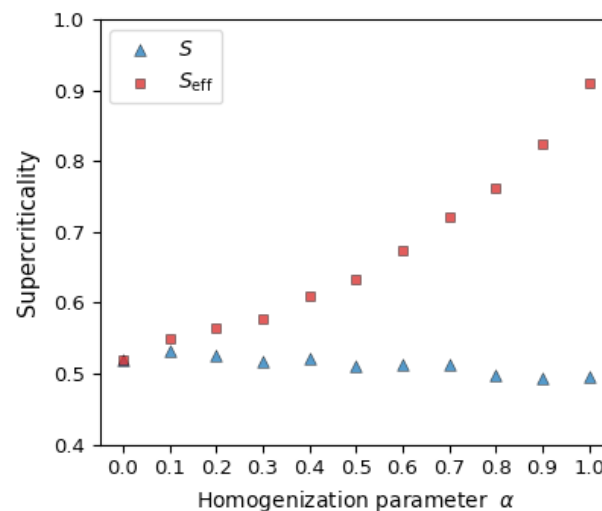


Figure 7. Supercriticality of Schneider and Walker (2006) computed using midlatitude (30°–60°) dry static stability (triangles) and effective static stability (squares) for a series of experiments in which α is varied from 0.0 to 1.0.

Considering that the tropopause height is not much different between the control and homogenization experiments, the occurrence of $S_{\text{eff}} \approx 0.5$ in the homogenization experiment implies that the vertical extent of baroclinic eddies is half that in the control experiment. However, inspection of the vertical distribution of eddy fluxes does not support this interpretation (not shown). Indeed, according to Schneider and Walker (2006) and Levine and Schneider (2015), such a subcritical regime is possible only when near-surface baroclinicity is weak, but the meridional temperature gradient becomes stronger when eddy-LH coupling is suppressed. We argue that the supercriticality in Eq. (18) does not necessarily constrain the equilibrated thermal structure of the moist extratropical atmosphere when LH is homogenized. It is of note that Jansen and Ferrari (2013) revisited Schneider’s supercriticality theory and demonstrated that the assumption of vertically uniform eddy diffusivity, employed in the original derivation, is not always valid. Their improved derivation accounts for the existence of supercritical states; however, it still does not explain the occurrence of the subcritical states found here. At present, the systematic variation of the meridional temperature gradient across the parameter sweep may be best understood qualitatively in terms of baroclinic adjustment.

b. Storm track activity

We next examine storm track activity. While linear moist baroclinic instability theory predicts that eddy-LH coupling invigorates individual baroclinic disturbances, it remains unclear whether the same expectation applies once the mean state is allowed to adjust. Figures 8a,b show the free-tropospheric integrals of EAPE ($\frac{c_p \gamma}{2} \overline{T'^2} \cos \phi$) and EKE ($\frac{1}{2} \overline{u'^2 + v'^2} \cos \phi$), where c_p is the specific heat at constant pressure, and γ is the stratification parameter defined as

$$\gamma = \frac{\bar{\theta}}{\bar{T}} \frac{R}{c_p p} \left(-\frac{\partial \theta}{\partial p} \right)^{-1}. \quad (19)$$

Here, R is the dry gas constant, and $-\frac{\partial \theta}{\partial p}$ is the hemispheric-mean dry static stability. The EKE and EAPE exhibit comparable magnitudes and similar latitudinal distributions, broadly consistent with the approximate equipartition found in idealized GCMs (Schneider and Walker 2006). Both components of eddy energy change little across the parameter sweep. Although EKE slightly decreases within 20°–40° as α varies from 1.0 to 0.0, EAPE exhibits only modest changes. To better focus on “transient” storm track activity, the same diagnostics are computed using only high-frequency (HF; high-pass-filtered with a 10-day period cutoff) eddies (Figs.

8c,d). In the control experiment, HF components of EKE and EAPE peak near 40°. As eddy-LH coupling is suppressed, neither HF EKE nor HF EAPE weakens; instead, they exhibit a slight poleward intensification, consistent with the poleward intensification of the EDJ with decreasing α (Fig. 5d).

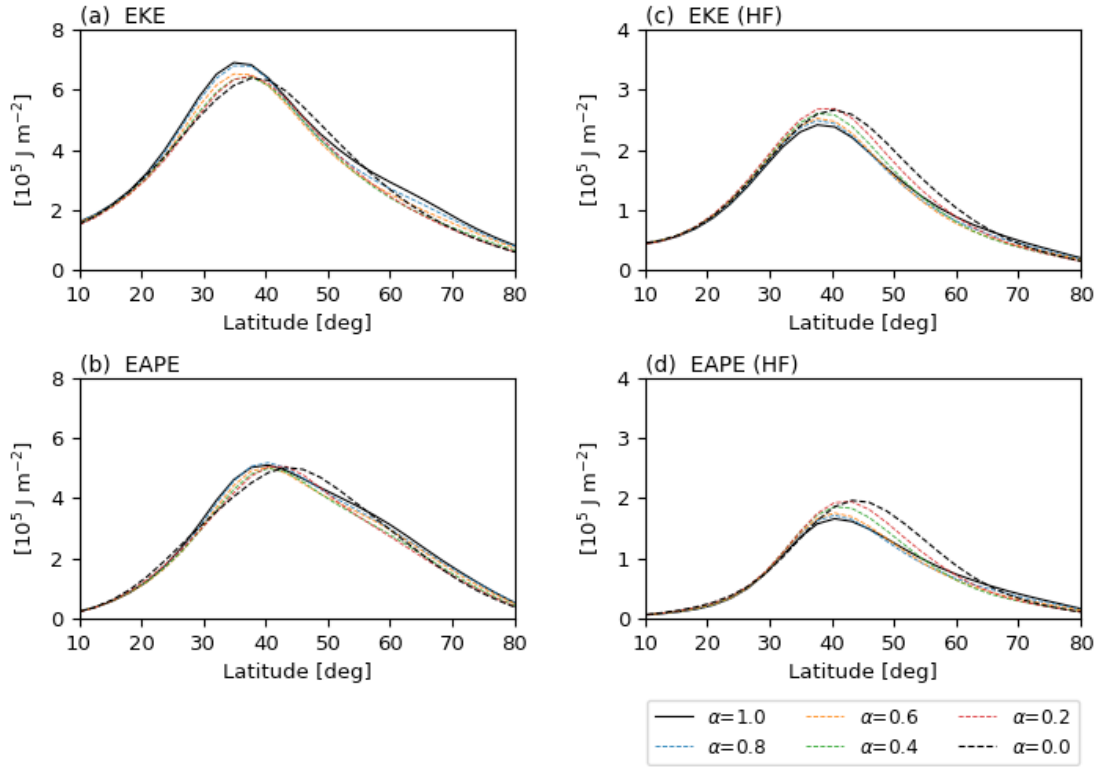


Figure 8. **a,b** Latitudinal profiles of vertically averaged (875–250 hPa) **a** EAPE and **b** EKE for a series of experiments in which α is varied from 0.0 to 1.0. The cosine latitude weighting is applied. **c,d** As in **a,b**, but only for HF eddies. Note that the y-axis range in **a,b** is twice that of **c,d**.

To explain this counterintuitive storm track response to the suppression of eddy-LH coupling, we next examine the budget of HF eddy energy based on the Lorenz energy cycle (Lorenz 1955; Peixóto and Oort 1974). Specifically, we consider the budget of the total eddy energy E , defined as the sum of EKE and EAPE. Although EAPE is not a direct measure of storm track activity, the advantage of the E equation over the EKE equation is that the internal conversion between EKE and EAPE cancels out, leaving only the external energy sources and sinks (Novak et al. 2020). Under the quasi-geostrophic approximation, the total eddy energy equation can be written as:

$$\frac{\partial \langle E \rangle}{\partial t} = \langle -c_p \gamma \overline{v' T'} \frac{\partial}{\partial \phi} \bar{T} \rangle + \langle \gamma \overline{T' Q'_{LH}} \rangle + \langle -\overline{u' v'} \frac{\partial}{\partial \phi} \bar{u} \rangle + \mathcal{R}, \quad (20)$$

where a is the radius of the earth; Q'_{LH} is the zonal deviation of LH rate per unit mass; $\mathcal{R}$ denotes the residual; angle brackets represent the hemispheric mass integration $\langle \cdot \rangle = -\int (\cdot) \frac{a^2}{g} \cos\phi d\lambda d\phi dp$. The 10-day high-pass filter is applied to all eddy quantities to focus on transient storm activity. In computing the stratification parameter γ in (Eq. 19), we do not replace the dry static stability with the effective static stability because the LH effect is explicitly represented by the second term on the right-hand side of Eq (20). Recalling Eq. (3), the eddy LH rate in our experiments is expressed as

$$Q'_{LH} = c_p \alpha \dot{T}'_{LH}. \quad (21)$$

The residual $\mathcal{R}$ includes diabatic generation of EAPE by processes other than LH, equatorial boundary energy inflows, and frictional dissipation, which are not considered explicitly.

The first three terms on the right-hand side of Eq. (20) represent baroclinic energy conversion from MAPE, diabatic generation of EAPE by LH, and barotropic energy conversion between eddy and mean kinetic energies, respectively. The left-hand side of Eq. (20) vanishes for a statistically equilibrated climate (i.e., eddies are saturated), and thus the energetic diagnostics alone cannot establish causality. Instead, they provide insight into the relative importance of the individual energetic processes that maintain storm track activity. Figure 9 compares the individual terms on the right-hand side of Eq. (20) among experiments with different α . For diagnostic purposes, we retain the latitudinal dependence by integrating only in the zonal and vertical directions (e.g., Novak et al. 2020). As α decreases from 1.0 to 0.0, baroclinic energy conversion undergoes a pronounced poleward strengthening (Fig. 9a), indicating enhanced extraction of MAPE by the eddies. Its peak magnitude is larger in the homogenization experiment than in the control experiment by a factor of 1.4, while the peak latitude changes only modestly across the parameter sweep.

In contrast, EAPE generation by LH decreases systematically with decreasing α and vanishes when $\alpha = 0.0$ (Fig. 9b). This behavior is expected because the LH contribution is determined by the covariance between T' and Q'_{LH} . In extratropical cyclones, latent heat is preferentially released in the warm sector where positive temperature anomalies are present (Browning et al. 1973), producing a positive covariance between T' and Q'_{LH} . Because Q'_{LH} is directly proportional to α in our experiments (Eq. 21), this contribution decreases systematically with decreasing α . Despite the contrasting responses of baroclinic energy conversion and EAPE generation by LH, their sum remains similar across the experiments (Fig. 9c). Thus, the loss of diabatic energy input is largely compensated by enhanced baroclinic

energy conversion, thereby maintaining storm-track activity despite the suppression of eddy–LH coupling.

The barotropic energy conversion, which is associated with the later stages of the baroclinic wave lifecycle, exhibits enhanced transfer of EKE to the mean flow at progressively higher latitudes on the equatorward flank of the storm track as α decreases (Fig. 9d). This behavior is likely related to changes in the relative positions of the storm track and the westerly jet across the parameter sweep. Nevertheless, the magnitude of this change is much smaller than those of the first two energetic terms and is therefore unlikely to play a major role in maintaining storm track activity.

The compensating increase in baroclinic energy conversion appears to be closely related to the adjustment of the mean thermal structure. To quantify this relationship, we assume that the HF eddy heat flux is local and approximately diffusive such that

$$\overline{v'T'} \approx -D \frac{\partial}{\partial \phi} \bar{T}, \quad (22)$$

where D is the eddy diffusivity. Rewriting the baroclinic energy conversion using Eq. (22) yields $\langle -c_p \gamma \overline{v'T'} \frac{\partial}{\partial \phi} \bar{T} \rangle \approx \langle c_p \gamma D \left(\frac{\partial}{\partial \phi} \bar{T} \right)^2 \rangle$. Assuming, for simplicity, that the eddy diffusivity has small vertical variation and remains approximately unchanged across the experiments, we obtain the following relation:

$$\langle -c_p \gamma \overline{v'T'} \frac{\partial}{\partial \phi} \bar{T} \rangle \propto \langle \gamma \left(\frac{\partial}{\partial \phi} \bar{T} \right)^2 \rangle, \quad (23)$$

which implies that the changes in dry baroclinic energy conversion across the experiments can be approximated by the product of the stratification parameter and the square of the meridional temperature gradient¹. Integrating over the latitudes where baroclinic energy conversion exhibits pronounced inter-experiment differences (40°–60°; Fig. 9a), we find that the baroclinic energy conversion increases by a factor of 1.94 from the control to the homogenization

¹ The diagnosed eddy diffusivity D modestly decreases with height in the control experiment. In the homogenization experiment, D decreases slightly in the lower troposphere and increases in the upper troposphere, becoming nearly vertically uniform. However, both the vertical structure and interexperimental differences in baroclinic energy conversion are dominated by $\gamma \left(\frac{\partial}{\partial \phi} \bar{T} \right)^2$, justifying the neglect of D in the approximated scaling in Eq. (23).

experiments, whereas the right-hand side of Eq. (23) increases by a factor of 1.92. This close agreement suggests that the adjustment of the mean thermal structure largely accounts for the compensating increase in baroclinic energy conversion. Because the dry static stability changes only weakly across the experiments, the stratification parameter γ also varies little. Consequently, the scaling is dominated by changes in the meridional temperature gradient. Indeed, omitting γ from the right-hand side of Eq. (23) alters the predicted increase only slightly, from 1.92 to 2.01, which remains close to the diagnosed ratio of 1.94. The compensating increase in baroclinic energy conversion is therefore primarily attributable to the enhanced meridional temperature gradient.

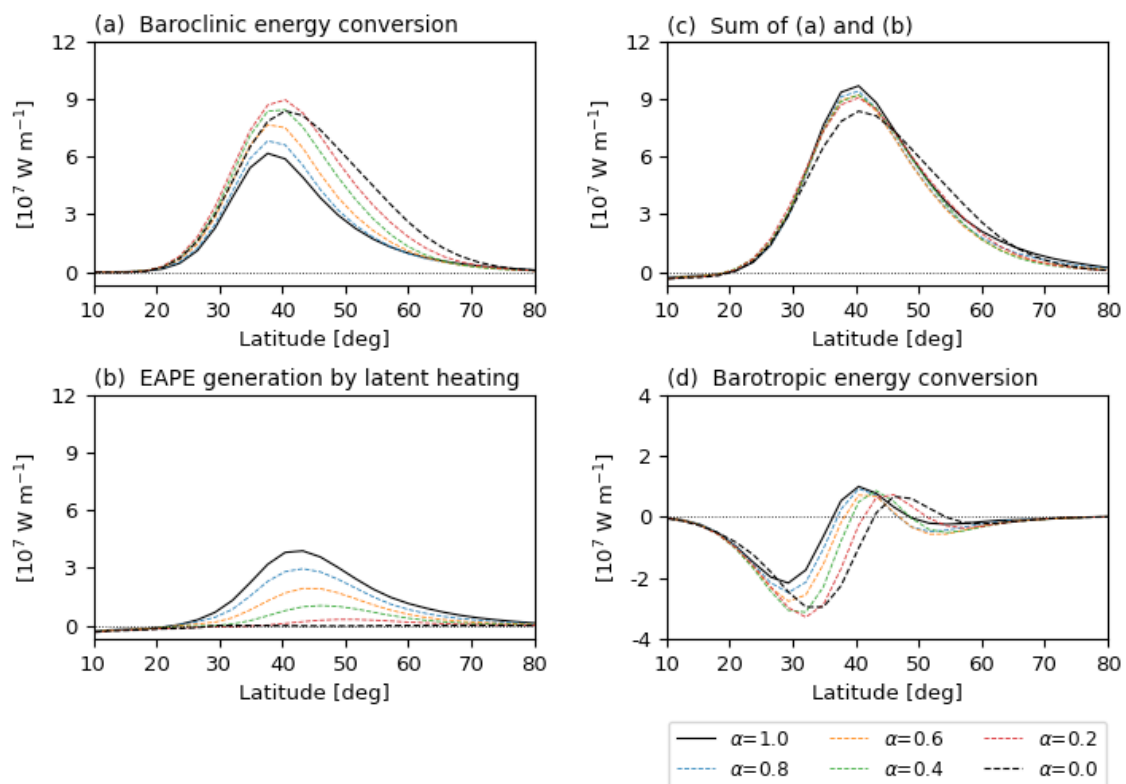


Figure 9. Latitudinal profiles of zonally and vertically integrated (875–250 hPa) energetic terms for HF eddies for a series of experiments in which α is varied from 0.0 to 1.0. **a** Baroclinic energy conversion, **b** EAPE generation by LH, **c** their sum, and **d** barotropic energy conversion. Note that the y-axis range of **d** is different from that in other panels.

It is also noteworthy that the right-hand side of Eq. (23) is similar (not identical to) the approximated MAPE of O’Gorman and Schneider (2008b) (see their Eq. 6). In moist idealized GCM simulations spanning a range of climates, O’Gorman and Schneider (2008b) showed that EKE scales approximately linearly with MAPE over the baroclinic zone. This scaling may explain the poleward strengthening of transient EKE found under eddy-LH decoupling (Fig.

8d), although the specific processes underlying the EKE response remain obscured from the steady state alone.

c. Eddy length scales

How does the characteristic scale of extratropical eddies respond to the suppression of eddy-LH coupling? In Earth-like dry atmospheres, the dominant scale of synoptic eddies is often comparable to that of the most unstable baroclinic waves, which in the Eady model is set by the Rossby deformation radius. This reflects weak eddy-eddy interactions and inefficient upscale energy transfer (Held 1999; Schneider 2004; O’Gorman and Schneider 2007). Since moist baroclinic instability preferentially destabilizes shorter waves, we examine whether the dominant eddy length scale changes with the suppression of eddy-LH coupling and whether its variation can be captured by the Rossby deformation radius.

Figure 10a shows the energy-containing zonal wavenumber ($\tilde{k}$) of upper-level eddy meridional wind. Following Kidston et al. (2010), $\tilde{k}$ is defined as

$$\tilde{k} = \frac{\sum_k k \cdot |\tilde{v}(k)|^2}{\sum_k |\tilde{v}(k)|^2}, \quad (24)$$

where $\tilde{v}(k)$ is the Fourier component of the eddy meridional wind at zonal wavenumber k . The $\tilde{k}$ decreases poleward from approximately $\tilde{k} = 9$ at 10° to $\tilde{k} = 2$ at 80° . The larger wavenumbers in the tropics may partly reflect the larger zonal circumference at low latitudes, as well as the prevalence of smaller-scale disturbances associated with moist convection in the absence of zonal asymmetry (Macdonald and Ming 2022). In the experiments with varying α , the midlatitude $\tilde{k}$ decreases systematically with decreasing α , implying longer wavelengths as eddy-LH coupling is suppressed. This response is qualitatively consistent with the prediction of moist baroclinic instability theory. Changes in $\tilde{k}$ are negligible in the tropics where the LH homogenization is not applied and at high latitudes where the influence of eddy-LH decoupling is limited by low moisture availability. Focusing on HF eddies, the decrease in midlatitude $\tilde{k}$ is more pronounced than for the total eddy (Fig. 10c). This stronger response is consistent with HF eddies being more representative of synoptic-scale disturbances, whose horizontal scales are more closely tied to the preferred length scales of baroclinic instability than those of low-frequency eddies. The lengthening of HF eddies is most pronounced on the poleward flank of the storm track (50° – 60°), coinciding with the region of the largest fractional increase in baroclinic energy conversion (Fig. 9a).

We next examine whether the variation in the energy-containing eddy length across the parameter sweep of α can be explained by the Rossby deformation radius. The energy-containing eddy length is defined as $\tilde{L} = 2\pi a \cos\phi / \tilde{k}$ and averaged over midlatitudes. The Rossby deformation radius is determined as

$$L_D = \frac{NH}{f}, \quad (25)$$

where $N = \left(\frac{g}{\theta} \frac{\partial \bar{\theta}}{\partial z}\right)^{1/2}$ is the Brunt–Väisälä frequency; g is the gravitational acceleration; the scale height H is somewhat arbitrary and set to 7 km. We further define the effective Rossby deformation radius by using the effective static stability in calculating the Brunt–Väisälä frequency as

$$L_{D,\text{eff}} = \frac{N_{\text{eff}}H}{f}, \quad (26)$$

where $N_{\text{eff}} = \left[\frac{g}{\theta} \left(\frac{\partial \bar{\theta}}{\partial z}\right)_{\text{eff}}\right]^{1/2}$. Figure 10b compares $\tilde{L}$ with L_D and $L_{D,\text{eff}}$. The L_D and $L_{D,\text{eff}}$ are rescaled by multiplicative constants determined from least-squares fits to $\tilde{L}$. The $\tilde{L}$ almost linearly decreases with α . This variation is reasonably well captured by the variation in $L_{D,\text{eff}}$. In contrast, L_D does not vary much and thus fails to reproduce the eddy lengthening with decreasing α . For HF eddies, the variation in $L_{D,\text{eff}}$ captures that in $\tilde{L}$ more closely (Fig. 10d), presumably reflecting their more direct association with the most unstable mode of linear baroclinic instability.

Although not shown, we also computed the Rhine scale $L_R = \sqrt{U'/\beta}$, where U' is the eddy velocity scale defined as $U' = \sqrt{2\text{EKE}}$ (Rhines 1975), and found that L_R change little across the parameter sweep of α . The weak sensitivity of the Rhines scale is expected from the modest changes in unfiltered EKE (Fig. 8a). This suggests that nonlinear turbulent upscale energy transfer, which is arrested at the Rhine scale, is not the primary control on the variation of the eddy length scale with the suppression of eddy-LH coupling.

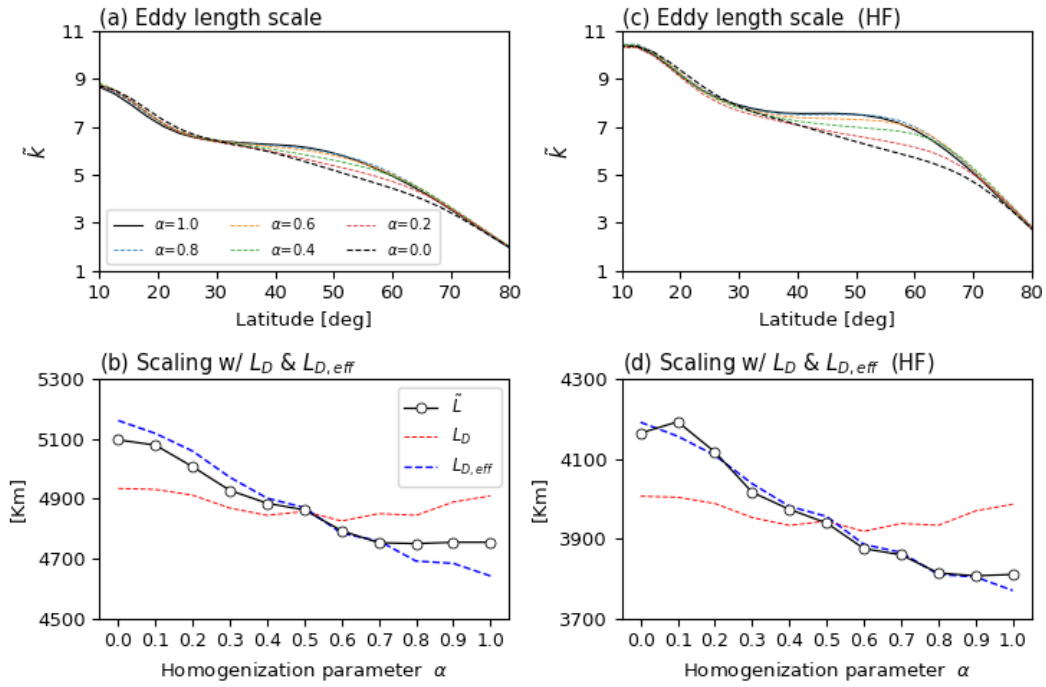


Figure 10. **a** Latitudinal profiles of dominant zonal wavenumber ($\tilde{k}$) of eddy meridional wind at 350 hPa for a series of experiments in which α is varied from 0.0 to 1.0. **b** Midlatitude (30°–60°) average of dominant eddy wavelength ($\tilde{L}$), Rossby deformation radius (L_D), and effective Rossby deformation radius ($L_{D,eff}$) at 350 hPa for a series of experiments in which α is varied from 0.0 to 1.0. L_D and $L_{D,eff}$ are rescaled by multiplicative factors obtained from least squares fit to $\tilde{L}$. **c,d** As in **a,b**, but for HF eddy.

To further illustrate the spectral signature of the increased eddy length scale under eddy-LH decoupling, Fig. 11a presents the wavenumber-phase speed power spectrum of upper-tropospheric eddy meridional wind. The spectra are obtained by applying a two-dimensional Fourier transform in time and longitude, followed by the Jacobian correction ($\Delta c/\Delta k$) to preserve the total spectral power (Randel and Held 1991). In the homogenization experiment, the spectral power shifts systematically toward lower zonal wavenumbers while exhibiting little change in ground-based phase speed. However, ground-based phase speed includes Doppler shifting by the zonal-mean flow, which undergoes substantial adjustment as eddy-LH coupling is suppressed. To remove this effect, Fig. 11b shows the corresponding spectrum as a function of the intrinsic phase speed, obtained by subtracting the zonal-mean zonal wind from the ground-based phase speed. In contrast to the ground-based phase speed, the intrinsic phase speed shifts systematically toward more westward propagation as the dominant spectral power moves toward lower zonal wavenumbers. This behavior is qualitatively consistent with the Rossby wave dispersion relation.

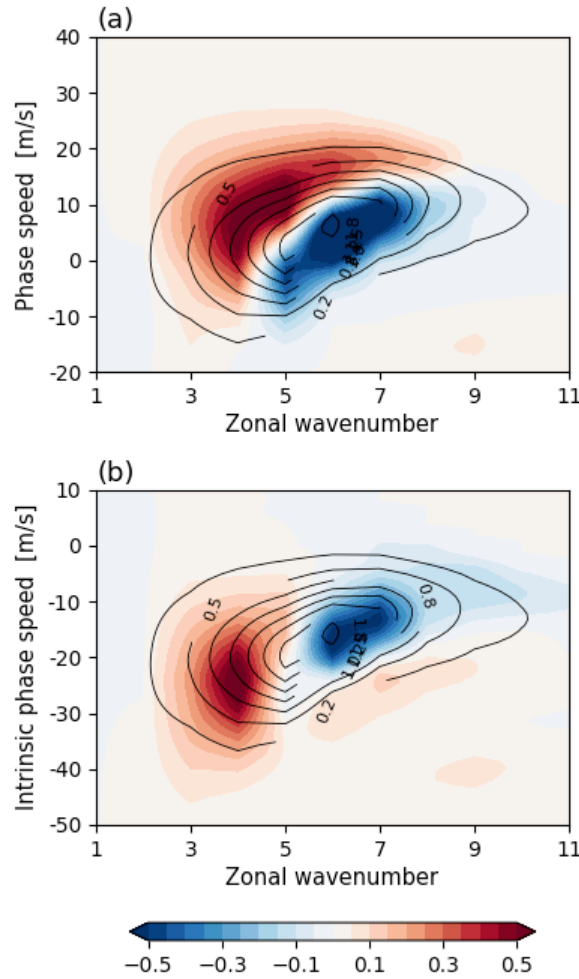


Figure 11. a Power spectral density of eddy meridional wind at 350 hPa as a function of zonal wavenumber and phase speed in the control experiment (contours, unit: $\text{m}^2 \text{s}^{-2} \Delta k^{-1} \Delta c^{-1}$) and its difference between the homogenization and control experiments ($\alpha=0.0$ minus $\alpha=1.0$; shading). Midlatitude ($30^\circ\text{--}60^\circ$) average is applied. **b** As in **a**, but as a function of zonal wavenumber and intrinsic phase speed.

5. Summary and discussion

We investigated how the equilibrated extratropical circulation responds to the suppression of eddy-LH coupling using an idealized moist GCM. To selectively isolate the role of eddy-LH coupling, we introduced a LH homogenization framework that suppresses only the zonally asymmetric component while preserving the zonal-mean LH. The novel advantage of this framework is that the strength of eddy-LH coupling can be systematically controlled through a tunable homogenization parameter, allowing systematic parameter-sweep experiments across the full range of coupling strengths. In this experimental design, the

effective static stability experienced by eddies varies continuously between the moist and dry limits as eddy-LH coupling is progressively suppressed.

Suppressing eddy-LH coupling leads to a systematic adjustment of the mean thermal structure, characterized by high-latitude cooling that strengthens the meridional temperature gradient in midlatitudes. In contrast, the tropospheric-mean dry static stability changes little. The strengthened temperature gradient is accompanied by a poleward shift and intensification of the EDJ. Across the parameter sweep experiments, the meridional temperature gradient increases approximately linearly with effective static stability. This covariation is qualitatively consistent with baroclinic adjustment in a moist atmosphere, although the supercriticality framework of Schneider and Walker (2006) does not quantitatively capture the response across the full range of coupling strengths.

Despite the suppression of eddy-LH coupling, the midlatitude storm track does not weaken but rather intensifies poleward. We found that storm-track activity is maintained through a redistribution of eddy energy sources. Although EAPE generation by LH shuts down when eddies are decoupled from LH, this energy loss is largely compensated by enhanced baroclinic energy conversion from MAPE. This compensation is closely linked to the adjusted mean thermal structure, particularly the strengthened meridional temperature gradient. This behavior contrasts with Lutsko et al. (2024) who found a weakening of EKE as moisture increases in a similar moist idealized GCM. The discrepancy arises because their experiments controlled the amounts of moisture, rather than isolating eddy-LH coupling. In their study, the EKE weakening was primarily associated with increased dry static stability which changes only weakly in the present experiments.

Despite the fully nonlinear nature of our simulations, the dominant extratropical eddy length scale increases systematically as eddies are decoupled from LH, consistent with the prediction from linear moist baroclinic instability. The variation in the dominant eddy length scale is successfully captured by the Rossby deformation radius when defined using the effective static stability. The variation is, however, not captured by the Rhine scale. This result suggests that upscale energy transfer remains sufficiently weak for the equilibrated eddy scale to be constrained by the effective Rossby deformation radius even when eddy-LH coupling is suppressed. Eddy lengthening is accompanied by a more westward intrinsic phase speed, qualitatively consistent with the Rossby wave dispersion relation.

An instructive comparison can be made with the recent study of Auestad et al. (2025), who suppressed eddy-LH coupling using a fundamentally different experimental design. They found that when the circulation is decoupled from LH, the westerly jet shifts poleward without changing its magnitude while midlatitude storms weaken, both contrasting with the present results. The contrasting responses may arise from the fundamental differences in the degree of thermodynamic adjustment permitted by the experimental designs. For example, the zonal-mean LH in our framework evolves interactively with the simulated circulation, whereas Auestad et al. (2025) prescribe a climatological three-dimensional LH field. Moreover, we allowed the slab ocean temperature to adjust through local atmosphere-ocean energy exchange, while their atmospheric GCM is forced by prescribed climatological sea surface temperature. Other differences in model configuration, including zonal asymmetry and clouds, may also contribute. Applying the present LH homogenization framework to GCMs of varying complexity will help assess the robustness of the circulation responses shown in this study.

The present study focuses exclusively on the equilibrated circulation and therefore cannot determine the causal processes through which the circulation adjusts to the suppression of eddy-LH coupling. Examining the transient evolution of the storm track, the westerly jet, and the mean thermal structure following an abrupt suppression of eddy-LH coupling would provide a useful way to identify the specific processes responsible for establishing the equilibrated circulation, including the poleward intensification of EKE and EDJ. This represents another important direction for future research.

A notable aspect of the present results is the concurrent eddy lengthening and poleward jet shift when eddies are decoupled from LH. A similar response has been reported under global warming, where changes in eddy length scale have been suggested to cause the poleward jet shift (Kidston et al. 2010, 2011; Rivière 2011). For example, Rivière (2011) argued that enhanced upper-tropospheric baroclinicity under global warming lengthens baroclinic eddies, promoting anticyclonic wave breaking and a poleward jet shift. Kidston et al. (2011) and Barnes and Hartmann (2011) also associated longer eddy wavelengths with a more poleward EDJ, although the latter considered a more general dynamical setting rather than global warming specifically. Whether the concurrent eddy lengthening and poleward jet shift identified here operate through one of these mechanisms remains an open question whose resolution would provide valuable insight into how eddy-LH coupling shapes jet dynamics.

The poleward shift of the midlatitude jet and the high-latitude cooling when extratropical eddies are decoupled from LH suggest that eddy-LH coupling anchors the jet farther equatorward while maintaining a warmer polar climate. If the dynamical importance of eddy-LH coupling increases in a warmer and moister climate, it may represent an additional mechanism contributing to future polar amplification, complementing the climate feedback processes that have traditionally been emphasized (Goosse et al. 2018). Likewise, eddy-LH coupling may influence the response of the midlatitude jet to climate change, alongside the widely discussed competition between tropical upper-tropospheric warming and polar amplification (Butler et al. 2010; Barnes and Screen 2015) and other proposed mechanisms (Shaw 2019). Extending the present framework across a wide range of climates may therefore help produce more reliable future climate projections and explain the spread in projected circulation responses among GCMs.

Acknowledgments. The constructive conversation with Lorenzo M. Polvani, Pu Lin, and Joonsuk Kang is appreciated. Computational resources were provided by the high-performance computing cluster at Boston College. Chanil Park was supported by faculty start-up funding provided by the University of Alaska Fairbanks. Seok-Woo Son was supported by the National Research Foundation of Korea (NRF) grant funded by the Korea government (MSIT) (2023R1A2C3005607).

Data Availability Statement. The moist idealized simulations were performed using the Isca framework publicly available at <https://github.com/execlim/isca>.

REFERENCES

- Auestad, H., A. Shibu, P. Ceppi, and T. Woollings, 2025: The Latent Heating Feedback on the Mid-Latitude Circulation. *Geophys. Res. Lett.*, **52**, <https://doi.org/10.1029/2025GL116437>.
- Baker, H. S., T. Woollings, and C. Mbengue, 2017: Eddy-driven jet sensitivity to diabatic heating in an idealized GCM. *J. Clim.*, **30**, 6413–6431, <https://doi.org/10.1175/JCLI-D-16-0864.1>.
- Balasubramanian, G., and M. K. Yau, 1996: Life Cycle of a Simulated Marine Cyclone: Energetics and PV Diagnostics. *J. Atmos. Sci.*, **53**, 639–653, [https://doi.org/https://doi.org/10.1175/1520-0469\(1996\)053<0639:TLCOAS>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1996)053<0639:TLCOAS>2.0.CO;2).
- Barnes, E. A., and D. L. Hartmann, 2011: Rossby Wave Scales, Propagation, and the Variability of Eddy-Driven Jets. *J. Atmos. Sci.*, **68**, 2893–2908, <https://doi.org/10.1175/JAS-D-11-039.1>.
- , and J. A. Screen, 2015: The impact of Arctic warming on the midlatitude jet-stream: Can it? Has it? Will it? *Wiley Interdiscip. Rev. Clim. Change*, **6**, 277–286, <https://doi.org/https://doi.org/10.1002/wcc.337>.
- Barry, L., G. C. Craig, and J. Thuburn, 2000: A GCM Investigation into the Nature of Baroclinic Adjustment. *J. Atmos. Sci.*, **57**, 1141–1155.
- Bembenek, E., D. N. Straub, and T. M. Merlis, 2020: Effects of Moisture in a Two-Layer Model of the Midlatitude Jet Stream. *J. Atmos. Sci.*, **77**, 131–147, <https://doi.org/10.1175/jas-d-19-0021.1>.
- Betts, A. K., and M. J. Miller, 1986: A new convective adjustment scheme. Part II: Single column tests using GATE wave, BOMEX, ATEX and arctic air-mass data sets. *Quarterly Journal of the Royal Meteorological Society*, **112**, 693–709, <https://doi.org/https://doi.org/10.1002/qj.49711247308>.
- Booth, J. F., S. Wang, and L. Polvani, 2013: Midlatitude storms in a moister world: Lessons from idealized baroclinic life cycle experiments. *Clim. Dyn.*, **41**, 787–802, <https://doi.org/10.1007/s00382-012-1472-3>.

761 ———, L. Polvani, P. A. O’Gorman, and S. Wang, 2015: Effective stability in a moist
762 baroclinic wave. *Atmospheric Science Letters*, **16**, 56–62,
763 <https://doi.org/10.1002/asl2.520>.

764 Brayshaw, D. J., B. Hoskins, and M. Blackburn, 2008: The storm-track response to idealized
765 SST perturbations in an aquaplanet GCM. *J. Atmos. Sci.*, **65**, 2842–2860,
766 <https://doi.org/https://doi.org/10.1175/2008JAS2657.1>.

767 Browning, K. A., M. E. Hardman, T. W. Harrold, and C. W. Pardoe, 1973: The structure of
768 rainbands within a mid-latitude depression. *Quarterly Journal of the Royal*
769 *Meteorological Society*, **99**, 215–231,
770 <https://doi.org/https://doi.org/10.1002/qj.49709942002>.

771 Butler, A. H., D. W. J. Thompson, and R. Heikes, 2010: The steady-state atmospheric
772 circulation response to climate change-like thermal forcings in a simple general
773 circulation model. *J. Clim.*, **23**, 3474–3496, <https://doi.org/10.1175/2010JCLI3228.1>.

774 Charney, J. G., 1947: The dynamics of long waves in a baroclinic westerly current. *J. Atmos.*
775 *Sci.*, **4**, 136–162, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(1947)004<0136:TDOLWI>2.0.CO;2)
776 [0469\(1947\)004<0136:TDOLWI>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1947)004<0136:TDOLWI>2.0.CO;2).

777 Davis, C. a., and K. a. Emanuel, 1991: Potential Vorticity Diagnostics of Cyclogenesis. *Mon.*
778 *Weather Rev.*, **119**, 1929–1953, [https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0493(1991)119<1929:PVDOC>2.0.CO;2)
779 [0493\(1991\)119<1929:PVDOC>2.0.CO;2](https://doi.org/10.1175/1520-0493(1991)119<1929:PVDOC>2.0.CO;2).

780 Dwyer, J. G., and P. A. O’Gorman, 2017: Moist Formulations of the Eliassen–Palm Flux and
781 Their Connection to the Surface Westerlies. *J. Atmos. Sci.*, **74**, 513–530,
782 <https://doi.org/10.1175/jas-d-16-0111.1>.

783 Eady, E. T., 1949: Long Waves and Cyclone Waves. *Tellus*, **1**, 33–52,
784 <https://doi.org/10.3402/tellusa.v1i3.8507>.

785 Edmon, H. J., B. J. Hoskins, and M. E. McIntyre, 1980: Eliassen–Palm Cross Sections for the
786 Troposphere. *J. Atmos. Sci.*, **37**, 2600–2616, [https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0469(1980)037<2600:EPCSFT>2.0.CO;2)
787 [0469\(1980\)037<2600:EPCSFT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<2600:EPCSFT>2.0.CO;2).

788 Emanuel, K. A., M. Fantini, and A. J. Thorpe, 1987: Baroclinic Instability in an Environment
789 of Small Stability to Slantwise Moist Convection. Part I: Two-Dimensional Models. *J.*

790 *Atmos. Sci.*, **44**, 1559–1573, <https://doi.org/https://doi.org/10.1175/1520->
791 0469(1987)044<1559:BIIAEO>2.0.CO;2.

792 ———, J. D. Neelin, and C. S. Bretherton, 1994: On large-scale circulations in convecting
793 atmospheres. *Quarterly Journal of the Royal Meteorological Society*, **120**, 1111–1143,
794 <https://doi.org/https://doi.org/10.1002/qj.49712051902>.

795 Frierson, D. M. W., 2007: The dynamics of idealized convection schemes and their effect on
796 the zonally averaged tropical circulation. *J. Atmos. Sci.*, **64**, 1959–1974,
797 <https://doi.org/10.1175/JAS3935.1>.

798 ———, 2008: Midlatitude static stability in simple and comprehensive general circulation
799 models. *J. Atmos. Sci.*, **65**, 1049–1062, <https://doi.org/10.1175/2007JAS2373.1>.

800 Frierson, D. M. W., I. M. Held, and P. Zurita-Gotor, 2006: A Gray-Radiation Aquaplanet
801 Moist GCM. Part I: Static Stability and Eddy Scale. *J. Atmos. Sci.*, **63**, 2548–2566,
802 <https://doi.org/https://doi.org/10.1175/JAS3753.1>.

803 Garfinkel, C. I., I. White, E. P. Gerber, M. Jucker, and M. Erez, 2020: The Building Blocks
804 of Northern Hemisphere Wintertime Stationary Waves. *J. Clim.*, **33**, 5611–5633,
805 <https://doi.org/https://doi.org/10.1175/JCLI-D-19-0181.1>.

806 Ghosh, S., O. Lachmy, and Y. Kaspi, 2024: The Role of Diabatic Heating in the Midlatitude
807 Atmospheric Circulation Response to Climate Change. *J. Clim.*, **37**, 2987–3009,
808 <https://doi.org/https://doi.org/10.1175/JCLI-D-23-0345.1>.

809 Goosse, H., and Coauthors, 2018: Quantifying climate feedbacks in polar regions. *Nature*
810 *Communications*, **9**, <https://doi.org/https://doi.org/10.1038/s41467-018-04173-0>.

811 Gutowski, W. J., L. E. Branscome, and D. A. Steward, 1992: Life Cycles of Moist Baroclinic
812 Eddies. *J. Atmos. Sci.*, **49**, 306–319, <https://doi.org/https://doi.org/10.1175/1520->
813 0469(1992)049<0306:LCOMBE>2.0.CO;2.

814 Held, I. M., 1999: The macroturbulence of the troposphere. *Tellus, Series A: Dynamic*
815 *Meteorology and Oceanography*, **51**, 59–70, <https://doi.org/10.1034/j.1600->
816 0870.1999.t01-1-00006.x.

817 Held, I. M., and A. Y. Hou, 1980: Nonlinear Axially Symmetric Circulations in a Nearly
818 Inviscid Atmosphere. *J. Atmos. Sci.*, **37**, 515–533,

819 [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(1980)037%3C0515:NASCIA%3E2.0.CO;2)
 820 [0469\(1980\)037%3C0515:NASCIA%3E2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1980)037%3C0515:NASCIA%3E2.0.CO;2).
 821 Held, I. M., and V. Larichev, 1996: A Scaling Theory for Horizontally Homogeneous,
 822 Baroclinically Unstable Flow on a Beta Plane. *J. Atmos. Sci.*, **53**, 946–952,
 823 [https://doi.org/https://doi.org/10.1175/1520-0469\(1996\)053<0946:ASTFHH>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1996)053<0946:ASTFHH>2.0.CO;2).
 824 Hoskins, B. J., M. E. McIntyre, and A. W. Robertson, 1985: On the use and significance of
 825 isentropic potential vorticity maps. *QUARTERLY JOURNAL OF THE ROYAL*
 826 *METEOROLOGICAL SOCIETY*, **111**, 877–946,
 827 <https://doi.org/10.2903/j.efsa.2016.4524/full>.
 828 Jansen, M., and R. Ferrari, 2013: The vertical structure of the eddy diffusivity and the
 829 equilibration of the extratropical atmosphere. *J. Atmos. Sci.*, **70**, 1456–1469,
 830 <https://doi.org/10.1175/JAS-D-12-086.1>.
 831 Jucker, M., and E. P. Gerber, 2017: Untangling the annual cycle of the tropical tropopause
 832 layer with an idealized moist model. *J. Clim.*, **30**, 7339–7358,
 833 <https://doi.org/https://doi.org/10.1175/JCLI-D-17-0127.1>.
 834 Jukes, M. N., 2000: The Static Stability of the Midlatitude Troposphere: The Relevance of
 835 Moisture. *J. Atmos. Sci.*, **57**, 3050–3057, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(2000)057<3050:TSSOTM>2.0.CO;2)
 836 [0469\(2000\)057<3050:TSSOTM>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(2000)057<3050:TSSOTM>2.0.CO;2).
 837 Kidston, J., S. M. Dean, J. A. Renwick, and G. K. Vallis, 2010: A robust increase in the eddy
 838 length scale in the simulation of future climates. *Geophys. Res. Lett.*, **37**, 1–4,
 839 <https://doi.org/10.1029/2009GL041615>.
 840 Kidston, J., G. K. Vallis, S. M. Dean, and J. A. Renwick, 2011: Can the increase in the eddy
 841 length scale under global warming cause the poleward shift of the jet streams? *J. Clim.*,
 842 **24**, 3764–3780, <https://doi.org/10.1175/2010JCLI3738.1>.
 843 Kirshbaum, D. J., T. M. Merlis, J. R. Gyakum, and R. McTaggart-Cowan, 2018: Sensitivity
 844 of idealized moist baroclinic waves to environmental temperature and moisture content.
 845 *J. Atmos. Sci.*, **75**, 337–360, <https://doi.org/10.1175/JAS-D-17-0188.1>.
 846 Kohl, M., and P. A. O’Gorman, 2022: The Diabatic Rossby Vortex: Growth Rate, Length
 847 Scale, and the Wave–Vortex Transition. *J. Atmos. Sci.*, **79**, 2739–2755,
 848 <https://doi.org/10.1175/JAS-D-22-0022.1>.

849 Lachmy, O., 2022: The Relation Between the Latitudinal Shifts of Midlatitude Diabatic
 850 Heating, Eddy Heat Flux, and the Eddy-Driven Jet in CMIP6 Models. *Journal of*
 851 *Geophysical Research: Atmospheres*, **127**, <https://doi.org/10.1029/2022JD036556>.
 852 ———, and Y. Kaspi, 2020: The Role of Diabatic Heating in Ferrel Cell Dynamics. *Geophys.*
 853 *Res. Lett.*, **47**, <https://doi.org/10.1029/2020GL090619>.
 854 Lee, S., and H. Kim, 2003: The Dynamical Relationship between Subtropical and Eddy-
 855 Driven Jets. *J. Atmos. Sci.*, **60**, 1490–1503, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/10.1175/1520-0469(2003)060<1490:TDRBSA>2.0.CO;2)
 856 [0469\(2003\)060<1490:TDRBSA>2.0.CO;2](https://doi.org/10.1175/1520-0469(2003)060<1490:TDRBSA>2.0.CO;2).
 857 Levine, X. J., and T. Schneider, 2015: Baroclinic Eddies and the extent of the Hadley
 858 circulation: An idealized GCM study. *J. Atmos. Sci.*, **72**, 2744–2761,
 859 <https://doi.org/10.1175/JAS-D-14-0152.1>.
 860 Lorenz, E. N., 1955: Available Potential Energy and the Maintenance of the General
 861 Circulation. *Tellus A: Dynamic Meteorology and Oceanography*, **7**, 157–167,
 862 <https://doi.org/10.3402/tellusa.v7i2.8796>.
 863 Lutsko, N. J., and M. C. Hell, 2021: Moisture and the persistence of annular modes. *J. Atmos.*
 864 *Sci.*, **78**, 3951–3964, <https://doi.org/10.1175/JAS-D-21-0055.1>.
 865 ———, J. Martinez-Claros, and D. D. B. Koll, 2024: Atmospheric Moisture Decreases
 866 Midlatitude Eddy Kinetic Energy. *J. Atmos. Sci.*, **81**, 1817–1832,
 867 <https://doi.org/https://doi.org/10.1175/JAS-D-23-0226.1>.
 868 Macdonald, C. G., and A. Y. I. Ming, 2022: Tropical Intraseasonal Variability Response to
 869 Zonally Asymmetric Forcing in an Idealized Moist GCM. *J. Clim.*, **35**, 4479–4501,
 870 <https://doi.org/10.1175/JCLI-D-22-0344.1>.
 871 Mak, M., 1982: On Moist Quasi-Geostrophic Baroclinic Instability. *J. Atmos. Sci.*, **39**, 2028–
 872 2037, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(1982)039<2028:OMQGBI>2.0.CO;2)
 873 [0469\(1982\)039<2028:OMQGBI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1982)039<2028:OMQGBI>2.0.CO;2).
 874 ———, 1994: Cyclogenesis in a conditionally unstable moist baroclinic atmosphere. *Tellus A*,
 875 **46**, 14–33, [https://doi.org/https://doi.org/10.1034/j.1600-](https://doi.org/https://doi.org/10.1034/j.1600-0870.1994.00003.x)
 876 [0870.1994.00003.x](https://doi.org/10.1034/j.1600-0870.1994.00003.x)DigitalObjectIdentifier(DOI).

877 Merlis, T. M., T. Schneider, S. Bordoni, and I. Eisenman, 2013: Hadley circulation response
878 to orbital precession. Part II: Subtropical continent. *J. Clim.*, **26**, 754–771,
879 <https://doi.org/10.1175/JCLI-D-12-00149.1>.

880 Mlawer, E. J., S. J. Taubman, P. D. Brown, M. J. Iacono, and S. A. Clough, 1997: Radiative
881 transfer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the
882 longwave. *Journal of Geophysical Research Atmospheres*, **102**, 16663–16682,
883 <https://doi.org/https://doi.org/10.1029/97JD00237>.

884 Moore, R. W., and M. T. Montgomery, 2004: Reexamining the dynamics of short-scale,
885 diabatic rossby waves and their role in midlatitude moist cyclogenesis. *J. Atmos. Sci.*,
886 **61**, 754–768, [https://doi.org/10.1175/1520-0469\(2004\)061<0754:RTDOSD>2.0.CO;2](https://doi.org/10.1175/1520-0469(2004)061<0754:RTDOSD>2.0.CO;2).

887 Nielsen-Gammon, J. W., and D. Keyser, 2000: Effective Stratification for Pseudoadiabatic
888 Ascent. *Mon. Weather Rev.*, **128**, 3007–3010.

889 Novak, L., T. Schneider, and F. Ait-Chaalal, 2020: Midwinter suppression of storm tracks in
890 an idealized zonally symmetric setting. *J. Atmos. Sci.*, **77**, 297–313,
891 <https://doi.org/https://doi.org/10.1175/JAS-D-18-0353.1>.

892 O’Gorman, P. A., 2011: The effective static stability experienced by eddies in a moist
893 atmosphere. *J. Atmos. Sci.*, **68**, 75–90, <https://doi.org/10.1175/2010JAS3537.1>.

894 —, and T. Schneider, 2007: Recovery of atmospheric flow statistics in a general circulation
895 model without nonlinear eddy-eddy interactions. *Geophys. Res. Lett.*, **34**,
896 <https://doi.org/10.1029/2007GL031779>.

897 —, and —, 2008a: The hydrological cycle over a wide range of climates simulated with
898 an idealized GCM. *J. Clim.*, **21**, 3815–3832, <https://doi.org/10.1175/2007JCLI2065.1>.

899 —, and —, 2008b: Energy of midlatitude transient eddies in idealized simulations of
900 changed climates. *J. Clim.*, **21**, 5797–5806, <https://doi.org/10.1175/2008JCLI2099.1>.

901 —, and —, 2009: The physical basis for increases in precipitation extremes in
902 simulations of 21st-century climate change. *Proc. Natl. Acad. Sci. U. S. A.*, **106**, 14773–
903 14777, <https://doi.org/10.1073/pnas.0907610106>.

904 Park, C., S.-W. Son, Y. N. Takayabu, S.-H. Park, D.-H. Cha, and E. J. Cha, 2023: Role of
905 Midlatitude Baroclinic Condition in Heavy Rainfall Events Directly Induced by Tropical

906 Cyclones in South Korea. *Mon. Weather Rev.*, **151**, 3113–3132,
 907 <https://doi.org/https://doi.org/10.1175/MWR-D-23-0046.1>.

908 Peixóto, J. P., and A. H. Oort, 1974: The annual distribution of atmospheric energy on a
 909 planetary scale. *J. Geophys. Res.*, **79**, 2149–2159,
 910 <https://doi.org/10.1029/jc079i015p02149>.

911 Petterssen, S., and S. J. Smebye, 1971: On the development of extratropical cyclones.
 912 *Quarterly Journal of the Royal Meteorological Society*, **97**, 457–482,
 913 <https://doi.org/10.1002/qj.49709741407>.

914 Phillips, N. A., 1954: Energy Transformations and Meridional Circulations associated with
 915 simple Baroclinic Waves in a two-level, Quasi-geostrophic Model. *Tellus*, **6**, 273–286,
 916 <https://doi.org/https://doi.org/10.1111/j.2153-3490.1954.tb01123.x>.

917 Randel, W. J., and I. M. Held, 1991: Phase Speed Spectra of Transient Eddy Fluxes and
 918 Critical Layer Absorption. *J. Atmos. Sci.*, **48**, 688–697,
 919 [https://doi.org/https://doi.org/10.1175/1520-0469\(1991\)048<0688:PSSOTE>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1991)048<0688:PSSOTE>2.0.CO;2).

920 Rhines, P. B., 1975: Waves and turbulence on a beta-plane. *J. Fluid Mech*, **69**, 417–443,
 921 <https://doi.org/https://doi.org/10.1017/S0022112075001504>.

922 Riehl, H., and J. S. Malkus, 1958: On the heat balance in the equatorial trough zone. *A reprint*
 923 *of Geophysics*.

924 Rivière, G., 2011: A dynamical interpretation of the poleward shift of the jet streams in
 925 global warming scenarios. *J. Atmos. Sci.*, **68**, 1253–1272,
 926 <https://doi.org/10.1175/2011JAS3641.1>.

927 Robinson, W. A., 2000: A Baroclinic Mechanism for the Eddy Feedback on the Zonal Index.
 928 *J. Atmos. Sci.*, **57**, 415–422, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(2000)057<0415:ABMFTE>2.0.CO;2)
 929 [0469\(2000\)057<0415:ABMFTE>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(2000)057<0415:ABMFTE>2.0.CO;2).

930 Schneider, T., 2004: The Tropopause and the Thermal Stratification in the Extratropics of a
 931 Dry Atmosphere. *J. Atmos. Sci.*, **61**, 1317–1340,
 932 [https://doi.org/https://doi.org/10.1175/1520-0469\(2004\)061<1317:TTATTTS>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(2004)061<1317:TTATTTS>2.0.CO;2).

933 ———, and C. C. Walker, 2006: Self-Organization of Atmospheric Macroturbulence into
 934 Critical States of Weak Nonlinear Eddy-Eddy Interactions. *J. Atmos. Sci.*, **63**, 1569–
 935 1586, <https://doi.org/https://doi.org/10.1175/JAS3699.1>.

936 ———, and P. A. O’Gorman, 2008: Moist convection and the thermal stratification of the
 937 extratropical troposphere. *J. Atmos. Sci.*, **65**, 3571–3583,
 938 <https://doi.org/10.1175/2008JAS2652.1>.

939 ———, ———, and X. J. Levine, 2010: Water vapor and the dynamics of climate changes.
 940 *Reviews of Geophysics*, **48**, 1–22, <https://doi.org/10.1029/2009RG000302>.

941 Shaw, T. A., 2019: Mechanisms of Future Predicted Changes in the Zonal Mean Mid-
 942 Latitude Circulation. *Curr. Clim. Change Rep.*

943 Son, S.-W., and S. Lee, 2005: The Response of Westerly Jets to Thermal Driving in a
 944 Primitive Equation Model. *J. Atmos. Sci.*, **62**, 3741–3757,
 945 <https://doi.org/https://doi.org/10.1175/JAS3571.1>.

946 Stoelinga, M. T., 1996: A Potential Vorticity-Based Study of the Role of Diabatic Heating
 947 and Friction in a Numerically Simulated Baroclinic Cyclone. *Mon. Weather Rev.*, **124**,
 948 849–874, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0493(1996)124<0849:APVBSO>2.0.CO;2)
 949 [0493\(1996\)124<0849:APVBSO>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0493(1996)124<0849:APVBSO>2.0.CO;2).

950 Stone, P., and L. Branscome, 1992: Diabatically Forced, Nearly Inviscid Eddy Regimes. *J.*
 951 *Atmos. Sci.*, **49**, 355–367, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(1992)049<0355:DFNIER>2.0.CO;2)
 952 [0469\(1992\)049<0355:DFNIER>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1992)049<0355:DFNIER>2.0.CO;2).

953 Stone, P. H., 1978: Baroclinic adjustment. *J. Atmos. Sci.*, **35**, 561–571,
 954 [https://doi.org/https://doi.org/10.1175/1520-0469\(1978\)035<0561:BA>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1978)035<0561:BA>2.0.CO;2).

955 Tamarin, T., and Y. Kaspi, 2016: The Poleward Motion of Extratropical Cyclones from a
 956 Potential Vorticity Tendency Analysis. *J. Atmos. Sci.*, **73**, 1687–1707,
 957 <https://doi.org/10.1175/JAS-D-15-0168.1>.

958 Thompson, D. W. J., and T. Birner, 2012: On the linkages between the tropospheric
 959 isentropic slope and eddy fluxes of heat during northern hemisphere winter. *J. Atmos.*
 960 *Sci.*, **69**, 1811–1823, <https://doi.org/10.1175/JAS-D-11-0187.1>.

961 Thorncroft, C. D., B. J. Hoskins, and M. E. McIntyre, 1993: Two paradigms of baroclinic-
 962 wave life-cycle behaviour. *Quarterly Journal of the Royal Meteorological Society*, **119**,
 963 17–55, <https://doi.org/10.1002/qj.49711950903>.

964 Troen, I. B., and L. Mahrt, 1986: A simple model of the atmospheric boundary layer;
 965 sensitivity to surface evaporation. *Boundary. Layer. Meteorol.*, **37**, 129–148.

- 966 Vallis, G. K., and Coauthors, 2018: Isca, v1.0: A framework for the global modelling of the
967 atmospheres of Earth and other planets at varying levels of complexity. *Geosci. Model*
968 *Dev.*, **11**, 843–859, <https://doi.org/10.5194/gmd-11-843-2018>.
- 969 Wang, B., and A. Barcilon, 1986: Moist Stability of a Baroclinic Zonal Flow with
970 Conditionally Unstable Stratification. *J. Atmos. Sci.*, **43**, 705–719,
971 [https://doi.org/https://doi.org/10.1175/1520-0469\(1986\)043<0705:MSOABZ>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1986)043<0705:MSOABZ>2.0.CO;2).
- 972 Wernli, H., and S. L. Gray, 2024: The importance of diabatic processes for the dynamics of
973 synoptic-scale extratropical weather systems – a review. *Weather and Climate*
974 *Dynamics*, **5**, 1299–1408, <https://doi.org/10.5194/wcd-5-1299-2024>.
- 975 Whitaker, J. S., and C. A. Davis, 1994: Cyclogenesis in a Saturated Environment. *J. Atmos.*
976 *Sci.*, **51**, 889–908, [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0469(1994)051<0889:CIASE>2.0.CO;2)
977 [0469\(1994\)051<0889:CIASE>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0469(1994)051<0889:CIASE>2.0.CO;2).
- 978 White, I. P., O. Lachmy, and N. Harnik, 2024: Influence of a local diabatic heating source on
979 the midlatitude circulation. *Quarterly Journal of the Royal Meteorological Society*, **150**,
980 5167–5187, <https://doi.org/10.1002/qj.4863>.
- 981 Wu, Y., and O. Pauluis, 2014: Midlatitude tropopause and low-level moisture. *J. Atmos. Sci.*,
982 **71**, 1187–1200, <https://doi.org/10.1175/JAS-D-13-0154.1>.
- 983 Xia, X., and E. K. M. Chang, 2014: Diabatic damping of zonal index variations. *J. Atmos.*
984 *Sci.*, **71**, 3090–3105, <https://doi.org/https://doi.org/10.1175/JAS-D-13-0292.1>.
- 985 Yang, G., and T. Li, 2023: Moist Baroclinic Instability along the Subtropical Mei-Yu Front.
986 *J. Clim.*, **36**, 805–822, <https://doi.org/10.1175/JCLI-D-22-0140.1>.
- 987 Zhao, Y., C. Park, and A. S.-W. Son, 2023: Importance of Diabatic Heating for the Eastward-
988 Moving Heavy Rainfall Events along the Yangtze River, China. *J. Atmos. Sci.*, **80**, 151–
989 165, <https://doi.org/https://doi.org/10.1175/JAS-D-21-0321.1>.
- 990 Zhou, S., and P. H. Stone, 1993a: The Role of Large-Scale Eddies in the Climate
991 Equilibrium. Part I: Fixed Static Stability. *J. Clim.*, **6**, 985–1001,
992 [https://doi.org/https://doi.org/10.1175/1520-0442\(1993\)006<0985:TROLSE>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0442(1993)006<0985:TROLSE>2.0.CO;2).
- 993 ———, and ———, 1993b: The Role of Large-Scale Eddies in the Climate Equilibrium. Part II:
994 Variable Static Stability. *J. Clim.*, **6**, 1871–1881,
995 [https://doi.org/https://doi.org/10.1175/1520-0442\(1993\)006<1871:TROLSE>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0442(1993)006<1871:TROLSE>2.0.CO;2).

996 Zurita-Gotor, P., 2005: Updraft/Downdraft Constraints for Moist Baroclinic Modes and Their
997 Implications for the Short-Wave Cutoff and Maximum Growth Rate. *J. Atmos. Sci.*, **62**,
998 4450–4458, [https://doi.org/https://doi.org/10.1175/JAS3630.1](https://doi.org/10.1175/JAS3630.1).

999