

RESEARCH ARTICLE

An assessment of radiative impacts of CO₂ on baroclinic instability using idealized life cycles

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Email: ahmadig@ut.ac.ir**Abstract**

Interaction between CO₂ and atmospheric radiation has a significant part in changing horizontal and vertical temperature distributions, through which it can affect the midlatitude atmospheric dynamics. Baroclinic instability, which is the energy source of large-scale eddy formation in midlatitudes, depends on meridional and vertical eddy fluxes of heat. While several studies have been devoted to the impact of CO₂ concentration present in the atmosphere on baroclinic eddies in climate time-scales, the current work aims to determine the corresponding impact over a period of 10 to 20 days relevant to the duration of baroclinic life cycles when the flow is deterministically predictable. To this end, the radiative parametrization scheme called RRTMG (Rapid Radiative Transfer Model for General Circulation Models) has been coupled with the Diabatic Contour-Advection Semi-Lagrangian (DCASL) dynamical core. Except for parametrization of radiation as well as surface processes to represent radiative interaction with the surface, no other physics parametrizations are included in the model. The idealized life-cycle experiments are carried out for both the predominantly anticyclonic (LC1) and cyclonic (LC2) Rossby wave breaking with the same initialization but five different CO₂ concentrations (0, 250, 500, 750 and 1,000 ppmv). The impacts of different CO₂ concentrations on eddy growth, mean flow and eddy-mean flow interaction are discussed. Results show that the presence of CO₂ has a damping effect on instability in the baroclinic growth stage, amounting to about 7% and 4% reduction in the peak value of domain area-average eddy kinetic energy for LC1 and LC2, respectively. In addition, increase in CO₂ concentration has a small negative impact on the growth rate, vertical eddy propagation and jet intensification.

KEYWORDSbaroclinic instability, CO₂, DCASL, radiative impact, radiative scheme, RRTMG

1 | INTRODUCTION

Interaction between greenhouse gases and atmospheric radiation plays a significant role in changing horizontal and vertical temperature gradients, through which radiation can affect the midlatitude atmospheric dynamics

(Hall *et al.*, 1994; Yin, 2005; Chang *et al.*, 2012; Yuval and Kaspi, 2016). Baroclinic instability, which is the mechanism of large-scale eddy formation in the midlatitudes, depends on poleward and vertical eddy temperature fluxes. In addition, the conversion of zonal into eddy available potential energy is related to a positive correlation between

poleward temperature flux and the basic-state poleward temperature gradient, which leads to generation of eddy kinetic energy and helps to sustain the growing eddy (Lee and Yoo, 2014).

Recent climatological studies have indicated that global warming due to increase in greenhouse gas concentrations causes the poleward shift of eddy-driven jet and storm tracks (Thompson *et al.*, 2000; Kushner *et al.*, 2001; Yin, 2005; Lorenz and DeWeaver, 2007; Barnes and Hartmann, 2011; Barnes and Polvani, 2013; Harvey *et al.*, 2014). Furthermore, a series of climate simulations, from an energetic point of view, has revealed that radiative forcing of greenhouse gases influences eddy–mean flow interaction in midlatitudes (Hernández-Deckers and von Storch, 2010; Yuval and Kaspi, 2016; Chemke, 2017). Wu *et al.* (2013) investigated the transient atmospheric circulation response in a seasonal time-scale to sudden doubling of CO₂ concentration by employing the National Center for Atmospheric Research Community Atmosphere Model version 3 (NCAR-CAM3) coupled to a Slab Ocean Model (SOM). While analysing the characteristics of eddies such as the phase speed and length-scale, their study was silent on the changes in the growth rate. Ceppi and Shepherd (2017) assessed changes in position and speed of the jet induced by CO₂ radiative forcing and climate feedbacks using CAM4 and ECMWF model HAMburg version 6 (ECHAM6: Stevens *et al.*, 2013) climate models. Climate feedbacks were imposed by surface albedo, cloud and water vapour. They suggested that uncertainty in radiative feedbacks causes different baroclinicity responses to CO₂ forcing. Chemke and Polvani (2019) surveyed the impact of an abrupt increase in CO₂ concentration on widening of the tropical belt through several mechanisms including changes in surface temperature, tropopause height, eddy phase speed and static stability. They showed that an increase in subtropical static stability (a decrease in temperature lapse rate), due to increase of CO₂ concentrations, is accompanied by poleward shifts of the Hadley cell edge and eddy momentum flux, thereby altering subtropical baroclinicity.

It is worth mentioning that previous studies have almost exclusively focused on climatological behaviour of atmospheric flow against different CO₂ concentrations. In other words, a majority of prior researches have simulated interaction between CO₂ radiative forcing and large-scale dynamics over climate time-scales, assessing the impacts of CO₂ enhancement on the mean flow characteristics. A challenging issue, which arises in this subject, is examining CO₂ radiative effects during the life cycle of eddies associated with baroclinic development. There are key questions that have not yet been discussed properly in the literature, for instance, how increase in CO₂ concentration may affect the growth and decay of large-scale eddies

in midlatitudes, whether increase of CO₂ concentration might change baroclinicity within a few days, and to what extent the response might depend on the type of life cycle. A further question is whether energy conversions during the life cycle of an eddy and thus the strength of eddies are sensitive to different amounts of CO₂ concentration. Also, it is not clear how an increase of CO₂ concentration acts on the mean flow and eddy–mean flow interaction in the time span of a life cycle. In addition, since the previous research was almost exclusively focused on qualitative results using complex climate models (e.g. Barnes and Polvani, 2013; Harvey *et al.*, 2014; Zappa and Shepherd, 2017), the main aim of this article is to give quantitative assessments on the questions raised above in a much simpler set-up. In this aspect, our work is in the spirit of developing a hierarchy of models to gain understanding of complex issues related to the atmospheric circulation (Vallis *et al.*, 2018).

To fill the aforementioned gaps in the literature, in the current study experiments have been performed under increasing CO₂ levels for sixteen days, which covers the main stages of baroclinic instability (Thorncroft *et al.*, 1993) and formation of synoptic-scale eddies in the midlatitudes. The purpose of this quantitative study is to assess the response of baroclinic development to the CO₂ radiative impacts in midlatitudes. Given that the baroclinic life-cycle problem is inherently related to the dynamics of storm tracks, the experiments and analyses provided may also help us in understanding the response of the extratropical storm tracks to climate change. To carry out the experiments, the Rapid Radiative Transfer Model for General Circulation Models (RRTMG) parametrization scheme has been coupled with the Diabatic Contour-Advection Semi-Lagrangian dynamical core called DCASL (Mohebalhojeh *et al.*, 2016; Laghaeizadeh *et al.*, 2020). To complete the coupling in terms of radiative interaction with the ground, a representation of the surface processes adopted from ECHAM6 has been used. The analytical formulation developed by Polvani and Esler (2007) is used to set up the initial conditions and carry out the life-cycle experiments with five different CO₂ concentrations. For a more complete picture of baroclinic development, both LC1 and LC2 life cycles are investigated which differ mainly in the type of Rossby wave breaking in the upper troposphere: anticyclonic for LC1 and cyclonic for LC2. The radiative parametrization scheme RRTMG is used to compute the cooling rate at each time step of the DCASL to take into account the radiative impact on the dynamics. This is in contrast to using the Newtonian relaxation method, which was common in previous studies (similar to Park *et al.* (2013)). Therefore, the cooling (heating) rate for different amounts of CO₂ concentration can be estimated at each stage in the life cycle of eddies.

The article is structured as follows. Section 2 describes the set-up of the numerical model in detail. In Section 3, the radiative effects of CO₂ concentrations on baroclinic instability, large-scale eddy formation and growth as well as the eddy-mean flow interaction are determined by dynamical diagnostics. Finally, the conclusions are presented in Section 4.

2 | SETTING UP THE NUMERICAL MODEL

2.1 | The dynamical core

The DCASL dynamical core for spherical primitive equations was first developed by Mohebalhojeh *et al.* (2016) in the generalized vertical coordinate (ξ) using the prognostic variables $Q = (f + \zeta)/(1 + \tilde{h})$, $\tilde{h}$ and δ with f the Coriolis parameter, ζ the relative vertical vorticity, $\tilde{h}$ a pressure thickness defined by $\partial p / \partial \xi = \partial \bar{p} / \partial \xi (1 + \tilde{h})$, and δ the horizontal divergence. Here p is the pressure and $\bar{p}$ is a reference pressure, which varies only with vertical direction. The generalized vertical coordinate is taken as a function of the sigma (σ) and the potential temperature (θ), the hybrid $\sigma - \theta$, where σ is a function of pressure, surface pressure (p_s) and pressure at the top of the model (p_{top}). What distinguishes DCASL from the other algorithms is the simultaneous use of a Lagrangian contour representation and an Eulerian grid representation for Q , i.e. the variable representing vertical flow. Laghaeizadeh *et al.* (2020) further extended the dynamical core to include the $\sigma - p$ vertical coordinate (Eckerman, 2008) in both the Charney–Phillips and Lorenz grids. The DCASL algorithm for which the results are reported here is that based on the $\sigma - p$ vertical coordinate in the Charney–Phillips grid where the potential temperature is stored on the interface of the layers called half-integer levels. When $\sigma - p$ is used for the vertical coordinate, Q becomes essentially the same as absolute vertical vorticity. The $\sigma - p$ vertical coordinate has been chosen here because the analytical formulation of initial conditions can be translated to this coordinate with no approximation, while the use of $\sigma - \theta$ entails problems leading to generation of imbalance. For the $\sigma - p$ vertical coordinate, pressure at the interface of layers is given by $p_{l+1/2} = A_{l+1/2}p_0 + B_{l+1/2}p_s$ with $p_0 = 10^5$ Pa; details on the arrays A and B are given in Table S1 of the supporting information.

2.2 | Adding the radiative transfer scheme to the dynamical core

In general circulation models, as part of the physics parametrization called by the solution procedure for the

thermodynamic energy equation, through computing the radiative cooling rate the temperature field is changed, which leads to interaction with the dynamical core. In this study, RRTMG developed at Atmospheric Environment Research (AER) is employed to compute short-wave and long-wave radiative fluxes and cooling rates (Clough *et al.*, 2005). RRTMG uses the correlated-k distribution method (Lacis and Oinas, 1991; Fu and Liou, 1992) and a transmittance look-up table to achieve high accuracy and efficiency (Mlawer *et al.*, 1997; Iacono *et al.*, 2008). To avoid complications, the impacts of other physical processes in the atmosphere on the temperature distribution have been ignored and RRTMG has been coupled with the dynamical core.

The values of cooling rate per unit mass (Q) are determined on the full-integer levels by the calculation of the net radiation fluxes (F) on the half-integer levels according to

$$Q = -\frac{1}{\rho} \frac{dF}{dz} = g \frac{dF}{dp}, \quad (1)$$

where ρ is density, g is acceleration due to gravity and $F = F^\uparrow - F^\downarrow$; $F^\uparrow$ and $F^\downarrow$ are respectively the upward and downward radiative fluxes or irradiances. In the DCASL dynamical core using the $\sigma - p$ vertical coordinate with the Charney–Phillips grid, the radiative cooling rates impact the potential temperature on the half-integer levels such that

$$\left(\frac{\partial \theta}{\partial t} \right)_{\text{rad}, l+1/2} = c_p \left(\frac{(\partial T / \partial t)_{\text{rad}, l+1/2} (\delta p)_{l+1} + (\partial T / \partial t)_{\text{rad}, l} (\delta p)_l}{\prod_{l+1} (\delta p)_{l+1} + \prod_l (\delta p)_l} \right),$$

where $(\partial \theta / \partial t)_{\text{rad}, l+1/2}$ denotes the potential temperature tendency at the half-integer level $l + 1/2$ due to radiative heating and thus temperature tendencies $(\partial T / \partial t)_{\text{rad}}$ at neighbouring layers. Also, $(\delta p)_l = p_{l+1/2} - p_{l-1/2}$ is the pressure thickness of the layer l , and $\Pi \equiv c_p (p/p_0)^{R/c_p}$ is the Exner pressure in which c_p is the specific heat capacity at constant pressure, $p_0 = 10^5$ Pa and R is the specific gas constant for dry air as no moisture is assumed in the experiments.

2.3 | The numerical experiments

2.3.1 | Initial conditions

The construction of initial conditions for the two life cycles follows Polvani and Esler (2007) with changes required for the prognostic variables used in $\sigma - p$ vertical coordinate with Charney–Phillips staggered grid. Each life cycle

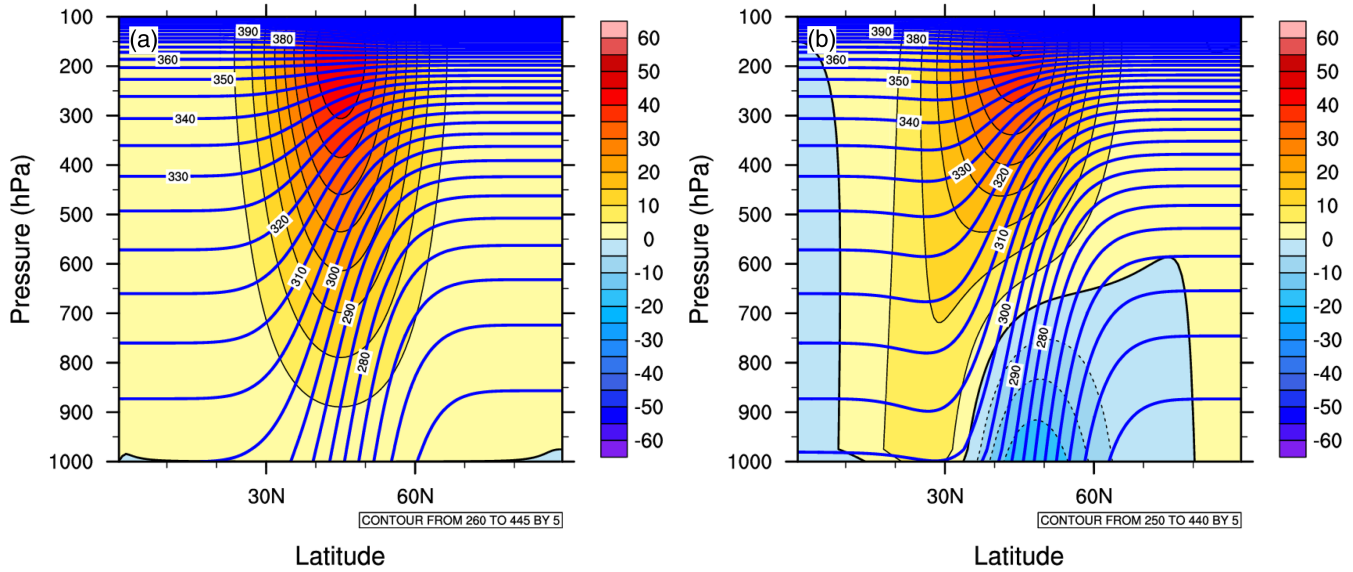


FIGURE 1 The vertical–meridional cross-section of the zonally averaged zonal wind (shaded, 5 m s^{-1} interval) and the potential temperature (thick blue solid lines, 5 K interval) at $t = 0$ for (a) LC1 and (b) LC2. The dark solid and dotted lines denote, respectively, the positive and negative values of the zonally averaged zonal wind [Colour figure can be viewed at wileyonlinelibrary.com]

is simulated for five cases with the same initialization but different CO_2 concentrations. The experiments are designed to alter the atmospheric optical depth by changing the values of CO_2 concentration to, respectively, 0, 250, 500, 750 and $1,000 \text{ ppmv}$. The CO_2 concentrations are considered uniformly distributed across the domain, both horizontally and vertically, throughout the time evolution. The initial conditions are defined in spherical coordinate (λ, φ, z) in which $\lambda \in [0, 2\pi]$ is longitude, $\varphi \in [-\pi/2, \pi/2]$ is latitude and $z \equiv H \log(p_0/p)$ is the log-pressure coordinate with H being the scale height.

The zonal winds u_1 for LC1 and u_2 for LC2 are initialized using the z coordinate with the scale height H taken to be equal to 7.5 km , such that

$$u_1(\varphi, z) = u_0 F(\varphi) \left[\frac{z}{z_T} e^{-[(z/z_T)^2 - 1]/2} \right],$$

$$u_2(\varphi, z) = u_1(\varphi, z) - U_s e^{-z/z_s} [\sin(2\varphi)]^2 \left[\frac{\varphi - \varphi_s}{\Delta_s} \right] \times e^{-[(\varphi - \varphi_s)/\Delta_s]^2}, \quad (2)$$

where $u_0 = 45 \text{ m s}^{-1}$ determines the maximum wind in the troposphere, $z_T = 13 \text{ km}$, $U_s = 45 \text{ m s}^{-1}$, $z_s = 10 \text{ km}$, $\varphi_s = 35^\circ$, $\Delta_s = 20^\circ$ and

$$F(\varphi) = \begin{cases} [\sin(\pi(\sin \varphi)^2)]^3 & \text{for } \varphi > 0, \\ 0 & \text{for } \varphi < 0. \end{cases}$$

All the five experiments use the same initial conditions for the zonal wind and the rest of variables presented in Appendix A. The number of grid points used in both meridional and zonal directions are 256, giving approximately grid lengths of 0.7° and 1.4° in respectively meridional and zonal directions. In the vertical direction, 55 layers are used covering from the surface to about 6.7 hPa pressure level.

Figure 1 shows the vertical–meridional cross-section of the zonally averaged zonal wind and the potential temperature at $t = 0$ (taken to be 21 March 2009, 1200 UTC) for both LC1 and LC2 in which the position of the upper-level jet is evident. These fields undergo development by the set-up of baroclinic instability.

To calculate cooling rate values, the vertical profiles of the pressure and the potential temperature are used to perform RRTMG. It should be noted that the radiative fluxes have been computed in aerosol-free and clear-sky conditions. In addition, the land surface scheme available in ECHAM6 has been used to account for the radiative interaction with the surface which is taken to be covered by water. No boundary-layer parametrization is carried out.

Since CO_2 does not have permanent (or electric) dipole moment, radiative transfer through pure rotational transitions is inactive. Nevertheless, it has vibrational motions due to its curved structure, causing disruptions in its linear symmetry, thereby producing oscillating dipole moment. Hence, CO_2 is a radiatively active gas in vibrational–rotational transitions (e.g. Petty, 2006). The CO_2 contribution to transitions and atmospheric cooling

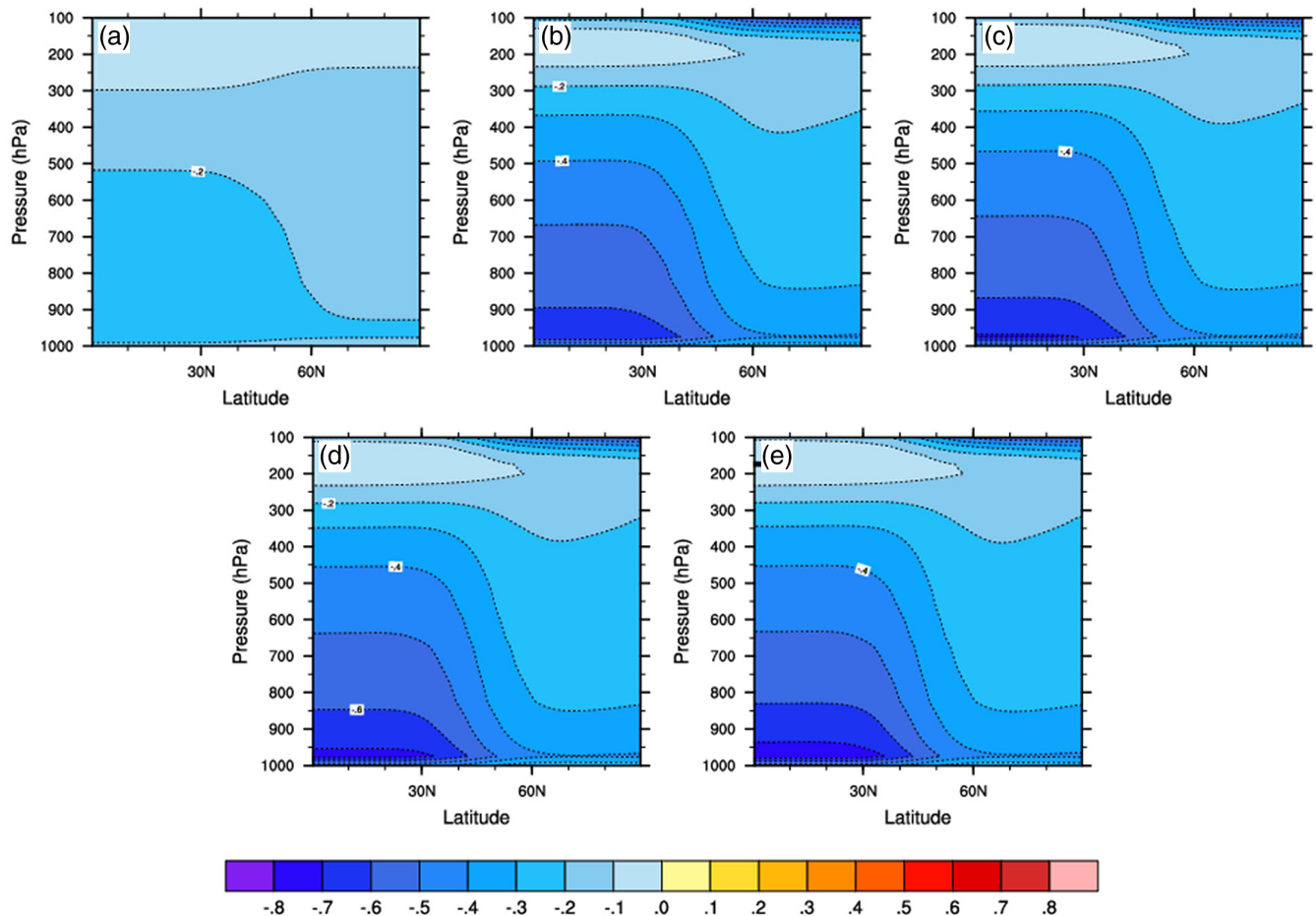


FIGURE 2 The vertical–meridional cross-section of the cooling rates (blue shaded, 0.1 K-day^{-1} interval) at $t = 0$ for LC1 and different CO_2 concentrations: (a) 0, (b) 250, (c) 500, (d) 750 and (e) 1,000 ppmv. Solid and dotted lines, respectively, show the positive and negative values [Colour figure can be viewed at wileyonlinelibrary.com]

(heating) is negligible for the short wavelength range (about 0.05 in all levels) and it transmits a considerable portion of solar radiation. However, it acts in the long wavelength range (IR band), because the atmosphere is opaque in the $14\text{--}16 \mu\text{m}$ band and the position of the $15 \mu\text{m}$ band is near the maximum Planck function emission for terrestrial temperatures.

Figure 2 depicts the vertical–meridional cross-section of the cooling rates, which are interpolated from the model's vertical coordinate to the standard pressure levels, at $t = 0$ for LC1 and different amounts of CO_2 (see Figure S1 for the corresponding results of LC2). As shown, the contribution of CO_2 to net radiative heating is rather small in the troposphere, because the emitted radiation is reabsorbed at nearby levels with nearly the same temperature. The cooling rates, however, exhibit a strong temperature dependency such that within the troposphere there is a well-defined Pole-to-Equator sloping gradient associated with the frontal zone represented by the potential temperature structure in Figure 1. The meridional gradient in radiative cooling is strongest near the surface

and increases for the higher amounts of CO_2 . The least radiative cooling is observed at the equatorial tropopause, where the temperature profile has a minimum value. The dominance of temperature effect in shaping the cooling rates is consistent with the cooling-to-space approximation for details of its validity; readers can refer to Jeevanjee and Fueglistaler (2020a). It is interesting to note the presence of more or less similar latitudinal–vertical structures in Figure 1a of Jeevanjee and Fueglistaler (2020b) for the zonal mean clear-sky long-wave heating rates computed using the reanalysis data of the European Centre for Medium-range Weather Forecasts (ECMWF) for June–August 2001. The markedly stronger cooling rates in their analysis are due to the effect of H_2O which is not considered here.

2.3.2 | Baroclinic instability

After defining the initial conditions, baroclinic instability is triggered by adding a height-independent perturbation

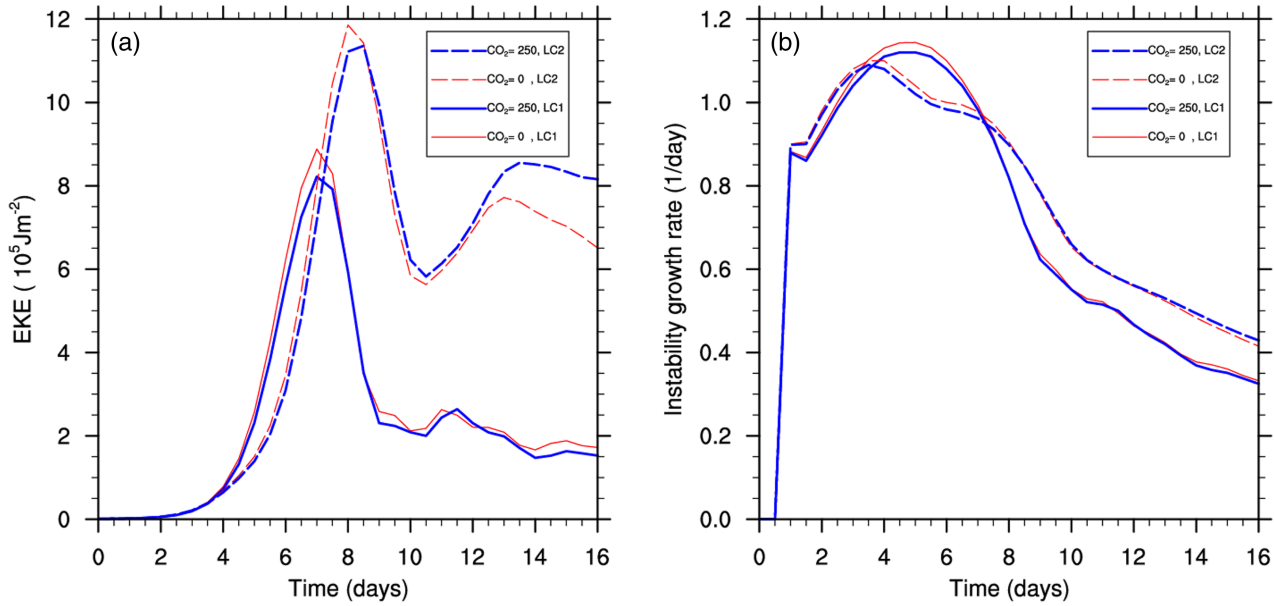


FIGURE 3 The time evolution of (a) EKE in unit of $10^5 \text{ J} \cdot \text{m}^{-2}$ and (b) growth rate in unit of day^{-1} for LC1 with zero (thin solid red line) and 250 ppmv (thick solid blue line) concentration of CO_2 , as well as LC2 with zero (thin dashed red line) and 250 ppmv (thick dashed blue line) concentration of CO_2 [Colour figure can be viewed at wileyonlinelibrary.com]

to the temperature field T' such that

$$T'(\lambda, \varphi) = \cos(6\lambda) [\text{sech}(6(\varphi - \hat{\varphi}))]^2, \quad (3)$$

where $\hat{\varphi} = 45^\circ \text{N}$. The perturbation is small enough for the imbalance generated to have negligible impact compared to the fastest growing normal mode that is rapidly excited leading to the formation of baroclinic instability.

3 | THE EFFECTS OF INCREASES IN CO_2 CONCENTRATION ON BAROCLINIC INSTABILITY

In this section, the development stages of baroclinic instability for different CO_2 concentrations are examined using some relevant dynamical diagnostics. While doing so, the radiative impacts of CO_2 on eddies, mean flow and eddy-mean flow interaction are studied during the baroclinic instability life cycle.

3.1 | Eddy kinetic energy and instability growth rate

The eddy kinetic energy is a domain area-average quantity defined by

$$EKE = \frac{\int_D ds \int_{p_{\text{top}}}^{p_s} \frac{1}{2} (u'^2 + v'^2) \frac{dp}{g}}{\int_D ds},$$

in which D denotes the horizontal domain, ds is the area element, u' and v' are respectively the deviation of zonal and meridional velocity components from their zonal averages. The horizontal averaging domain D is taken to be the Northern Hemisphere as in the experiments designed the Southern Hemisphere is baroclinically inactive. As mentioned earlier, EKE provides the basic information on the state of baroclinic instability and its life cycle for each experiment. Furthermore, the formation and evolution of instability can be studied using the growth rate given by $\sigma = [\ln(EKE(t)/EKE(0))]/t$, where $EKE(t)$ and $EKE(0)$ denote the eddy kinetic energies at time t and time zero, respectively.

Figure 3a shows the time evolution of EKE during the two life cycles for experiments with zero and 250 ppmv CO_2 concentration. The period of simulations is sixteen days and results are stored at 0.5 day intervals. As their generic characteristics, in both the life cycles, an early growth until around day 4 is followed by a stage of rapid growth associated with vertical Rossby-wave propagation and then a barotropic decay which is marked in LC1. In terms of the peak value of EKE, there are about 7.4% and 4.2% reductions in the presence of CO_2 for LC1 and LC2, respectively. While small compared to the reduction in EKE due to boundary-layer friction with values amounting to as high as 50% (Adamson *et al.*, 2006), the effect of CO_2 is nevertheless sign-definite and in the way of damping EKE during the growth phase. This can be seen by looking at the growth rates shown in Figure 3b as well as the differences in EKE between the experiments with and without

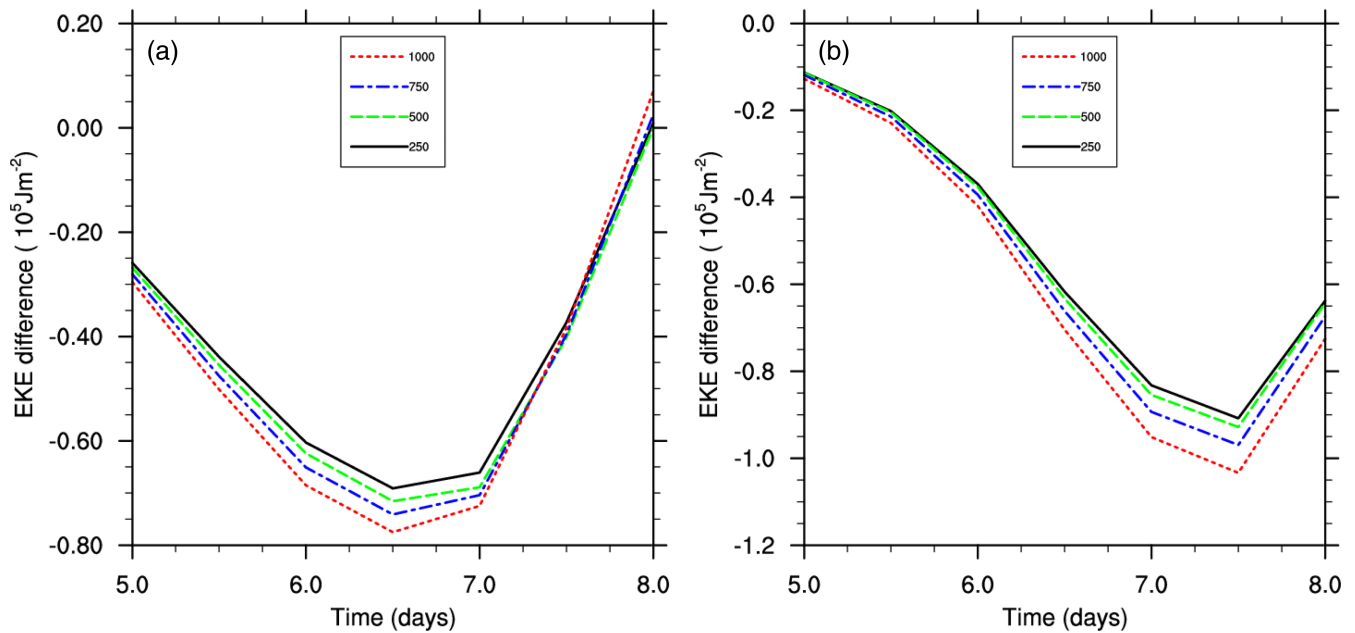


FIGURE 4 The time evolution of EKE difference in units of $10^5 \text{ J} \cdot \text{m}^{-2}$ for (a) LC1 and (b) LC2 during the growth phase from $t = 5$ days to $t = 8$ days. The differences are: $\text{EKE}_{250} - \text{EKE}_0$ (dark), $\text{EKE}_{500} - \text{EKE}_0$ (dashed green), $\text{EKE}_{750} - \text{EKE}_0$ (dash-dotted blue), and $\text{EKE}_{1000} - \text{EKE}_0$ (dotted red) [Colour figure can be viewed at wileyonlinelibrary.com]

TABLE 1 The maximum values of the domain area-average EKE ($10^5 \text{ J} \cdot \text{m}^{-2}$) and σ (day^{-1}) for different CO_2 concentrations and the two life cycles

	0 ppmv	250 ppmv	500 ppmv	750 ppmv	1000 ppmv
EKE (LC1)	8.881	8.220 (7.44)	8.192 (7.76)	8.177 (7.93)	8.156 (8.16)
EKE (LC2)	11.856	11.357 (4.21)	11.376 (4.05)	11.367 (4.12)	11.349 (4.28)
σ (LC1)	1.144	1.123 (1.84)	1.122 (1.92)	1.121 (2.01)	1.119 (2.19)
σ (LC2)	1.102	1.088 (1.27)	1.087 (1.36)	1.087 (1.36)	1.086 (1.45)

Note: The numbers in parentheses are the percentage reductions in the maximum EKE and growth rate values with respect to those of zero CO_2 experiments.

CO_2 (i.e. $\text{EKE}_{250-0} = \text{EKE}(\text{CO}_2 = 250) - \text{EKE}(\text{CO}_2 = 0)$ and the other differences with similar definitions) as shown in Figure 4 during days 5 and 8. It is also confirmed by the information provided in Table 1 on the maximum domain area-average EKE and σ for different experiments and the percentage relative reduction in maximum EKE and growth rate with respect to the zero CO_2 experiments.

3.2 | Radiative heating rates

Figure 5 displays the latitude–vertical cross-sections of the radiative heating (cooling) rates at the times $t = 5$ days during the growth stage and $t = 8$ days right after the peak of EKE for LC1 in the experiment without CO_2 (Figure 5a,d) and that with 250 ppmv CO_2 concentration. The corresponding results for LC2 are given in Figure S2. Near the surface, in the presence of CO_2 the

heating rates attain values largely different from the rest of the troposphere; for clarity, results are presented separately for the [100,850] hPa and [850,1000] hPa layers. Throughout the layer [100,850] hPa, CO_2 has a cooling effect with a distinct slope in midlatitudes due to the frontal zone and stronger cooling rates are observed at the lower latitudes of the lower troposphere. Comparing the results of days 5 and 8, it is apparent that both the midlatitude slope and Equator-to-Pole gradient in the cooling rate decrease during the time evolution by the heat transfer through baroclinic eddies. Near to the surface in Figure 5b for LC1, a narrow cooling region close to the northward edge of the baroclinic zone (see Figure 1) at 60° N can be seen that is sandwiched between two heating regions, a stronger in the south and a weaker in the north. The heating–cooling–heating structure is sharper and deeper at day 8 (Figure 5e) and can be understood in terms of the competing effects of cooling-to-space and the

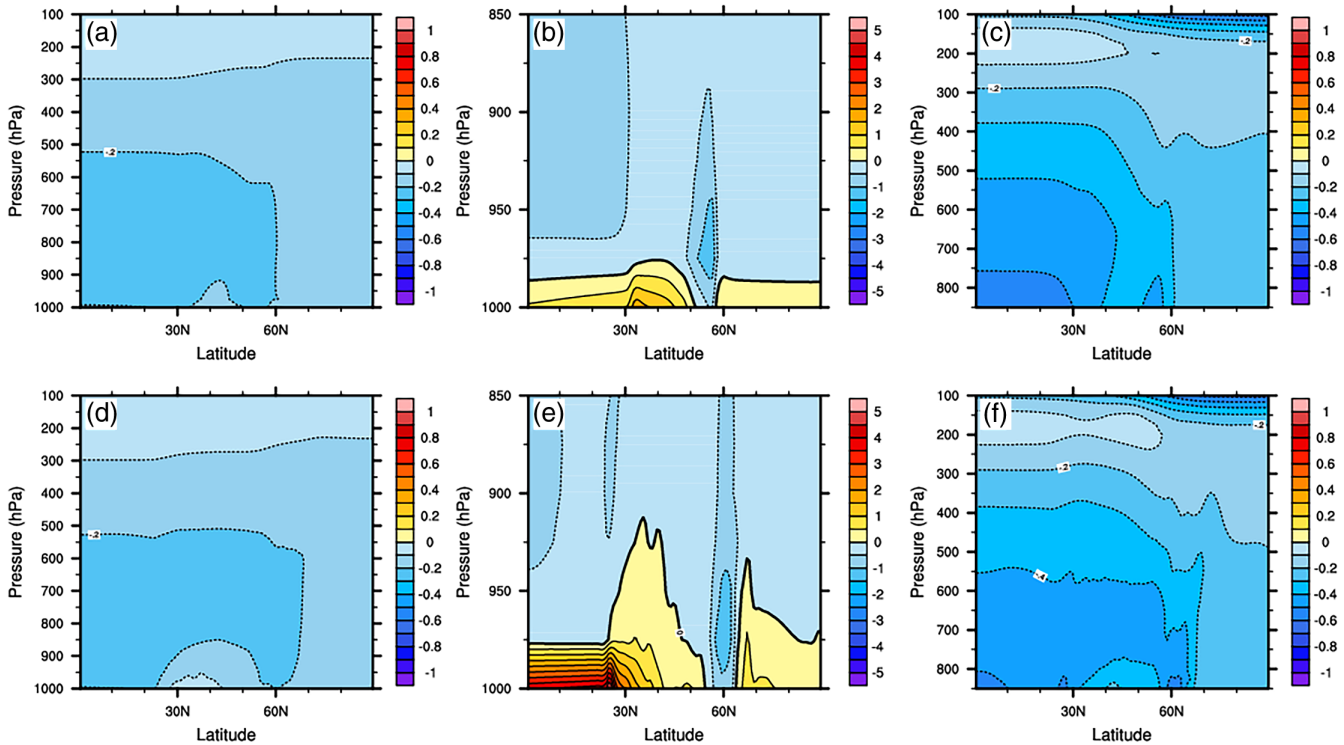


FIGURE 5 The vertical–meridional cross-sections of heating rates in $\text{K} \cdot \text{day}^{-1}$ for LC1 with CO_2 concentration of 0 ppmv (a,d) and 250 ppmv (b,c,e,f) at $t = 5$ days (a–c) and $t = 8$ days (d–f). For better presentation, the vertical interval 1,000 to 100 hPa has been separated to two subintervals of 1,000 to 850 hPa (b,e) and 850 to 100 hPa (c,f) [Colour figure can be viewed at wileyonlinelibrary.com]

surface heating induced by radiative exchange of the surface with upper layers. For LC2 (Figure S2e) which has a stronger surface baroclinic zone, the heating turns into a frontal-like structure close to 60°N . The radiative effect near to the surface is opposite that of the temperature advection.

3.3 | Mean flow

As mentioned before, the eddy kinetic energy is derived from the eddy available potential energy, which is dependent on the mean available potential energy. Since the meridional gradient of potential temperature is proportional to the mean available potential energy, it can affect eddy kinetic energy indirectly and at the same time responds to the radiative heating. To see this, shown in Figures 6 and 7 are the time–height sections of, respectively, LC1 and LC2 for $\Delta\bar{\theta}_{250} - \Delta\bar{\theta}_0$ and $\Delta\bar{\theta}_{1000} - \Delta\bar{\theta}_0$ where $\Delta\bar{\theta} = \bar{\theta}(30^\circ \text{N}) - \bar{\theta}(60^\circ \text{N})$. The quantity $\Delta\bar{\theta}$ is used here as a measure of temperature gradient across the baroclinic zone. Except for a small interval near to time $t = 6$ days, CO_2 across the layer [200, 1000] hPa in LC1 acts to reduce the meridional gradient of temperature, an effect which is especially stronger near the surface for 1,000 ppmv concentration (compare Figure 6a,c). In contrast

with LC1, in LC2 there is a sign change in the impact of CO_2 such that from day 4 onwards CO_2 acts to increase the surface baroclinicity (Figure 7a,c) and thus helps the maintenance of EKE (Figure 3a). When compared with the changes in EKE due to radiative impacts of CO_2 presented in Section 3.1, the contrasting behaviour of LC1 and LC2 in terms of changes in the meridional temperature gradient appears consistent with the findings of O’Gorman and Schneider (2008) and O’Gorman (2010) on a linear relation between EKE and mean available potential energy.

Since the reduction of mean meridional temperature gradient is associated with a decrease in the vertical wind shear (due to the thermal wind balance), so the presence of CO_2 in increasing values affects the zonally averaged zonal wind $\bar{u}$. Figure 8 shows $\bar{u}_{250} - \bar{u}_0$ and $\bar{u}_{1000} - \bar{u}_0$ at around the peak of EKE once the baroclinic instability is fully developed (i.e. day 7 for LC1 and day 8 for LC2). While the changes due to CO_2 are not generally sign-definite, associated with the zonal mean heating seen in Figure 5, the marked feature is the juxtaposition of the positive and negative values stretching from the surface to the upper troposphere near to 60°N , especially for LC2, in a way to increase (decrease) the cyclonic shear north of the jet in LC1 (LC2). Although not undertaken here, this feature can be quantitatively related to the corresponding structures discussed for the radiative heating in Figures 5 and S2

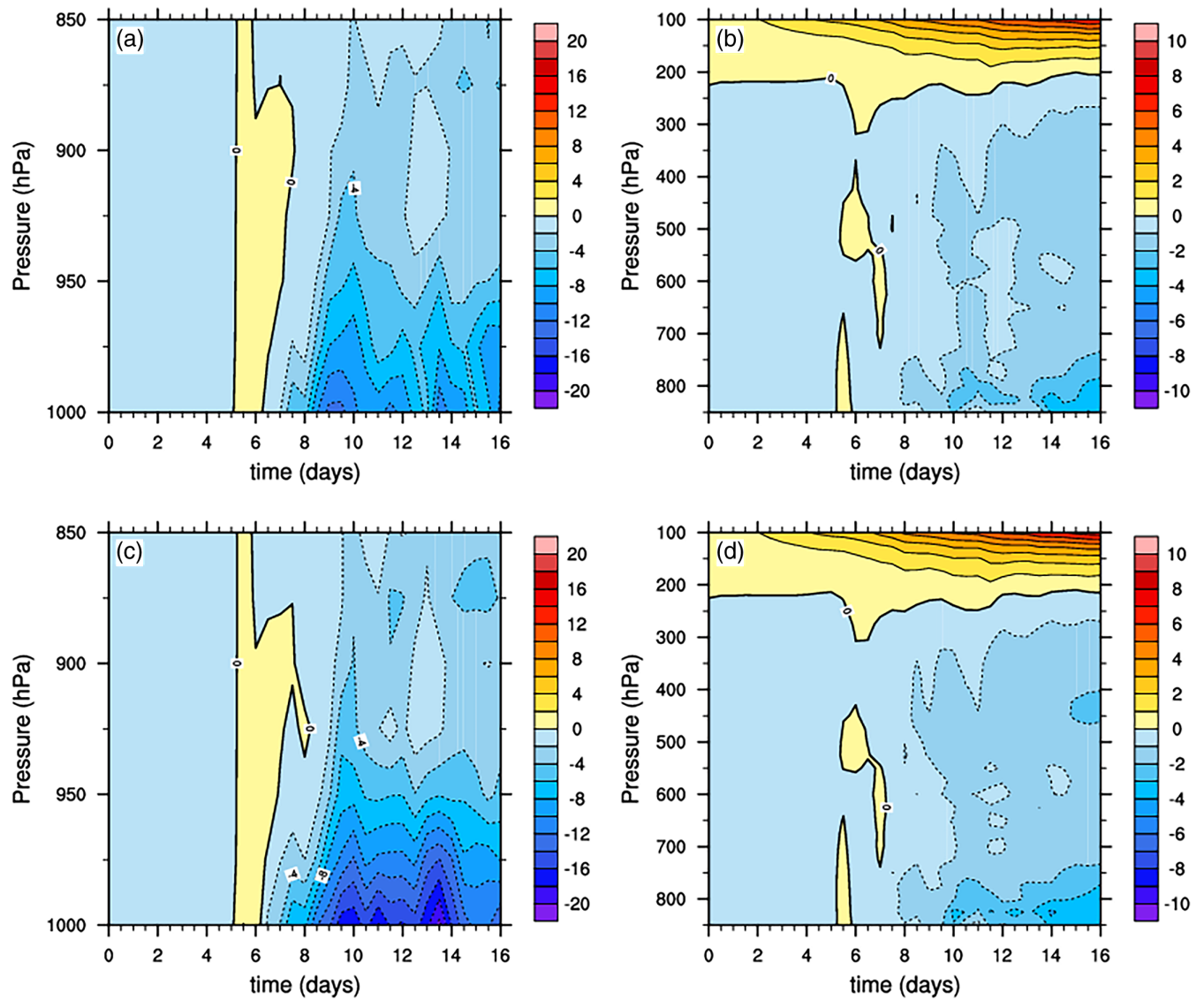


FIGURE 6 The time–height section of $\Delta\bar{\theta}_{250} - \Delta\bar{\theta}_0$ (a,b) and $\Delta\bar{\theta}_{1000} - \Delta\bar{\theta}_0$ (c,d) for LC1 where $\Delta\bar{\theta} = \bar{\theta}(30^\circ\text{N}) - \bar{\theta}(60^\circ\text{N})$. For better presentation, the vertical interval 1,000 to 100 hPa has been separated to two subintervals of 1,000 to 850 hPa (a,c) and 850 to 100 hPa (b,d) [Colour figure can be viewed at wileyonlinelibrary.com]

using the elliptic equation governing the relation between the zonal mean acceleration and the zonal mean diabatic forcing in the quasi-geostrophic model (see equation 3.5.7 in Andrews *et al.* (1987)). Any associated changes of the momentum transfer in the upper troposphere can be studied using the diagnostics presented in the next subsection.

3.4 | Eddy–mean flow interaction

In what has been presented so far, the effects of increase in CO_2 concentration on cooling rate, eddy activity and mean flow were investigated separately using some relevant quantities. In what follows, the radiative impacts of CO_2 on the eddy–mean flow interaction are presented

by calculating the Eliassen–Palm flux (EP flux) in each experiment. The EP flux is given by (Edmon *et al.*, 1980)

$$\mathbf{F} = (F^{(\varphi)}, F^{(p)}); \quad F^{(\varphi)} = -\overline{v'u'}, \quad F^{(p)} = f \frac{\overline{v'\theta'}}{\partial\theta/\partial p}, \quad (4)$$

where bar and prime denote the zonal mean and the corresponding perturbation, respectively. In addition, the superscripts (φ) and (p) denote the poleward and vertical directions.

The EP flux relation used in this study is based on the quasi-geostrophic equations (QG) on the β plane with the pressure vertical coordinate. As shown in Equation (4), the vertical ($F^{(p)}$) and meridional ($F^{(\varphi)}$) components of the EP flux represent the meridional temperature and

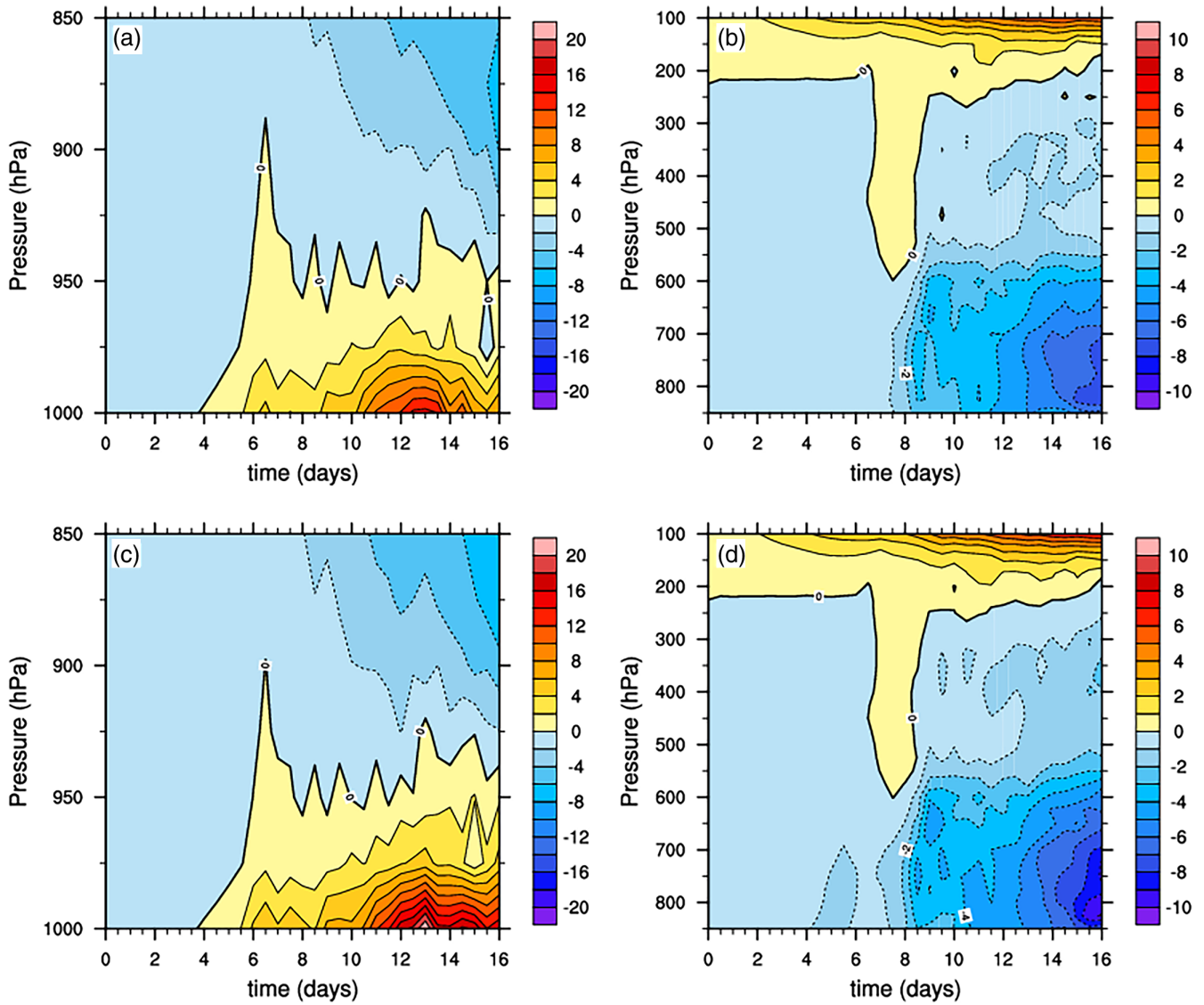


FIGURE 7 Same as Figure 6 but for LC2 [Colour figure can be viewed at wileyonlinelibrary.com]

momentum fluxes, respectively. Therefore, $F^{(p)}$ indicates the conversion of mean available potential energy to eddy available potential energy and $F^{(\varphi)}$ is associated with the conversion of eddy available potential energy to eddy kinetic energy. Furthermore, the EP flux divergence is a crucial quantity in displaying the eddy–mean flow interaction. This quantity is equivalent to the meridional eddy potential vorticity flux ($\overline{v'q'}$) in the QG theory, q' being the perturbation QG potential vorticity. The positive (negative) values of potential vorticity flux can cause the acceleration (deceleration) of zonal wind and/or generate a meridional circulation (Vallis, 2017). The EP flux divergence is given by (Edmon *et al.*, 1980):

$$\nabla \cdot \mathbf{F} = \overline{v'q'} = \frac{\partial}{\partial y} (-\overline{v'u'}) + \frac{\partial}{\partial p} \left(f \frac{\overline{v'\theta'}}{\partial \bar{\theta} / \partial p} \right). \quad (5)$$

Figures 9 and 10 show the zonally averaged zonal wind, the EP flux vector and its divergence at $t = 5$ days and $t = 8$ days for zero and 500 ppmv CO_2 concentration experiments of LC1 and LC2, respectively. For the graphical display in the latitude–pressure coordinates, the flux components $F^{(p)}$ and $F^{(\varphi)}$ are scaled with factors given in Appendix B. As the major features discussed in the analysis are not affected by the small-scale noise, no spatial smoothing operation has been applied to the model fields for the computation of EP diagnostics. The plots are restricted to the [100, 900] hPa layer, because for the 500 ppmv CO_2 concentration experiments $\partial \bar{\theta} / \partial p$ vanishes and changes sign in the [900, 1000] hPa layer (see Figure S3) leading to exceedingly large values of the vertical EP flux and its divergence. To fix ideas, in the description below, areas of flux divergence and flux convergence refer to regions where $\nabla \cdot \mathbf{F} > 0$ and $\nabla \cdot \mathbf{F} < 0$, respectively.

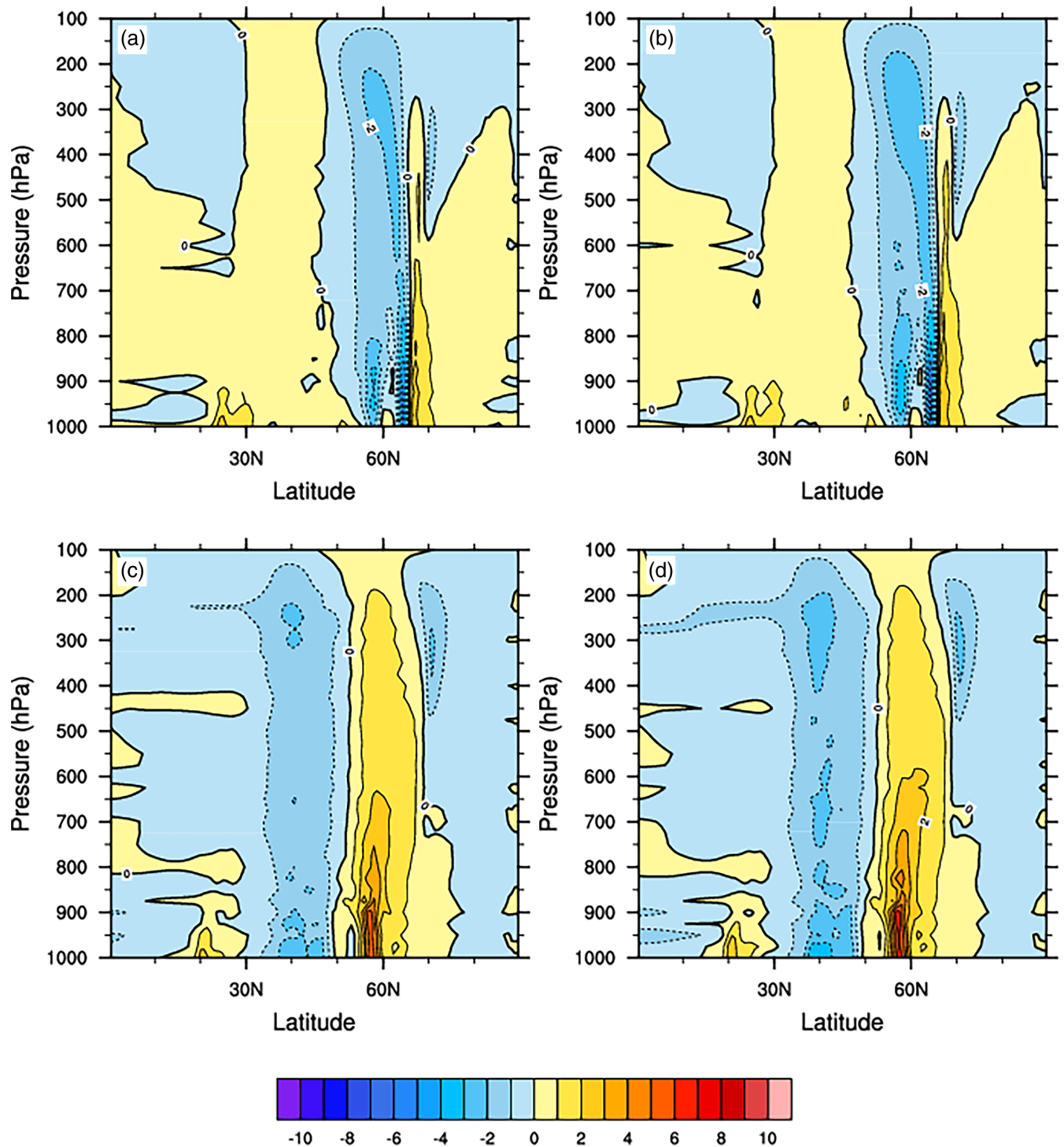


FIGURE 8 The vertical–meridional cross-sections of zonally averaged zonal wind differences ($\text{m}\cdot\text{s}^{-1}$) for $\bar{U}_{250-0}$ (a,c) and $\bar{U}_{1000-0}$ (b,d). The results are shown for LC1 (a,b) at $t = 7$ days and LC2 (c,d) at $t = 8$ days [Colour figure can be viewed at wileyonlinelibrary.com]

As shown in Figure 9a for LC1, at time $t = 5$ days which coincides with nearly the start of the rapid growth of EKE, one can identify a pair of the EP flux divergence/convergence in the lower troposphere around 45° latitude associated with which there is a distinct signature of vertical propagation. Being slightly weaker in the experiment

with 500 ppmv CO_2 concentration (Figure 9c), the EP flux convergence area is stretched upward near to the jet region in the upper troposphere. At time $t = 8$ days which is at the early stage of EKE decay, the main features in Figure 9b,d are the presence of a strong flux divergence between 45° and 60° latitudes as well as a weaker flux

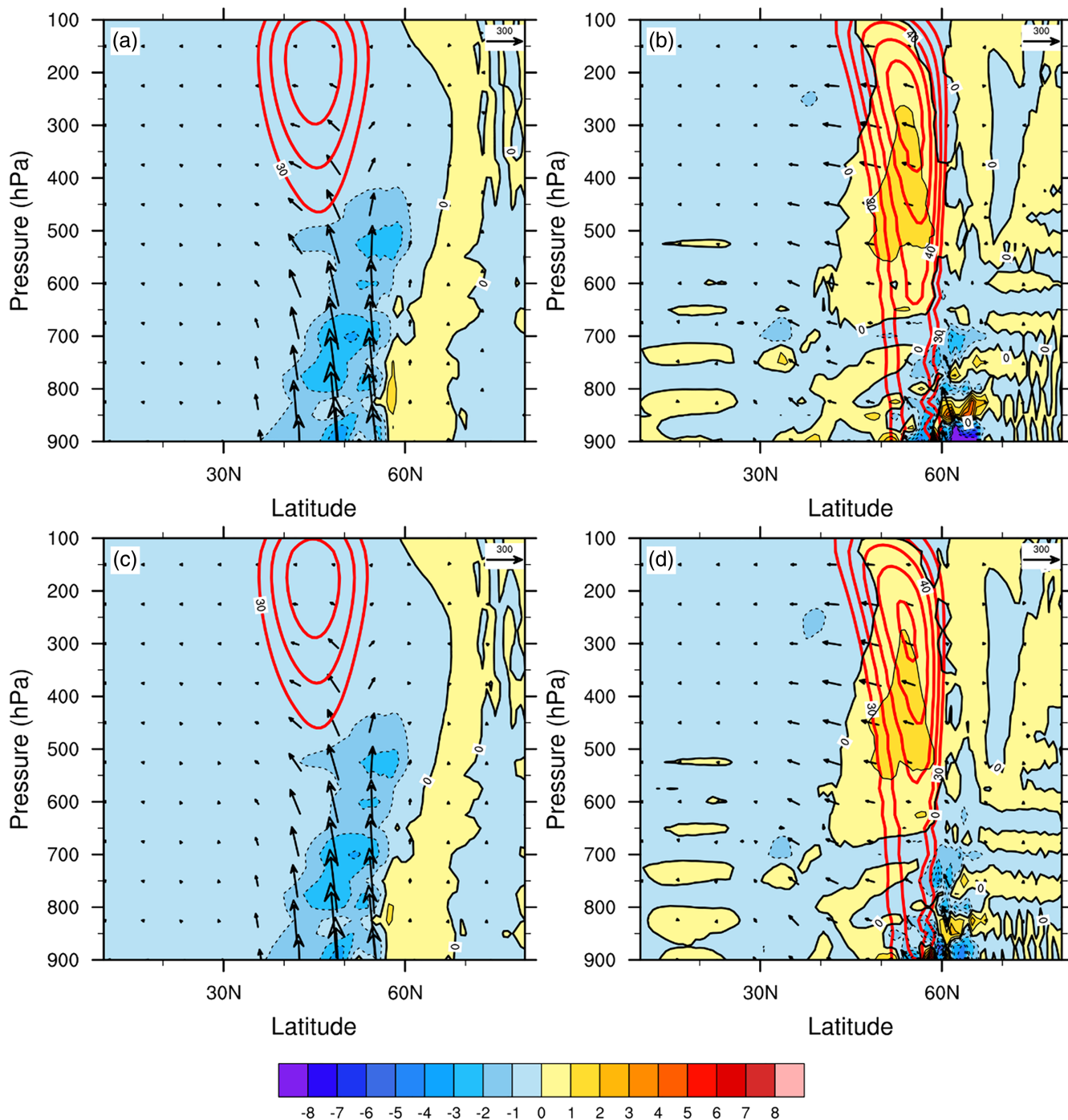


FIGURE 9 The zonally averaged zonal wind ($\geq 30 \text{ m} \cdot \text{s}^{-1}$ with $5 \text{ m} \cdot \text{s}^{-1}$ interval, thick solid red lines), the EP flux ($\text{kg} \cdot \text{m} \cdot \text{s}^{-2}$, black arrows), the EP flux divergence (shaded with $10^3 \text{ m} \cdot \text{s}^{-1} \cdot \text{day}^{-1}$ contour interval) for LC1 at day 5 (a,c) and day 8 (b,d) with zero CO_2 (a,b) and 500 ppmv CO_2 (c,d) concentrations. The solid and dashed lines denote, respectively, the divergence ($\nabla \cdot \mathbf{F} > 0$) and convergence ($\nabla \cdot \mathbf{F} < 0$) areas of EP flux [Colour figure can be viewed at wileyonlinelibrary.com]

convergence between 30° and 40° latitudes, both stretched vertically in the $[200, 600]$ hPa layer. What is striking is the extent to which the position of the jet and its downward extension leading to barotropization of the flow are connected to the EP flux divergence area. Theoretically, the vertical extension of divergence (convergence) areas to the

middle troposphere causes the acceleration (deceleration) of zonally averaged zonal wind and the subsequent downward extension of the upper-level jet stream in a manner consistent with midlatitude eddy-driven jet dynamics (Valis, 2017). The related difference fields shown in Figure S4 indicate that the main radiative impact of CO_2 is in making

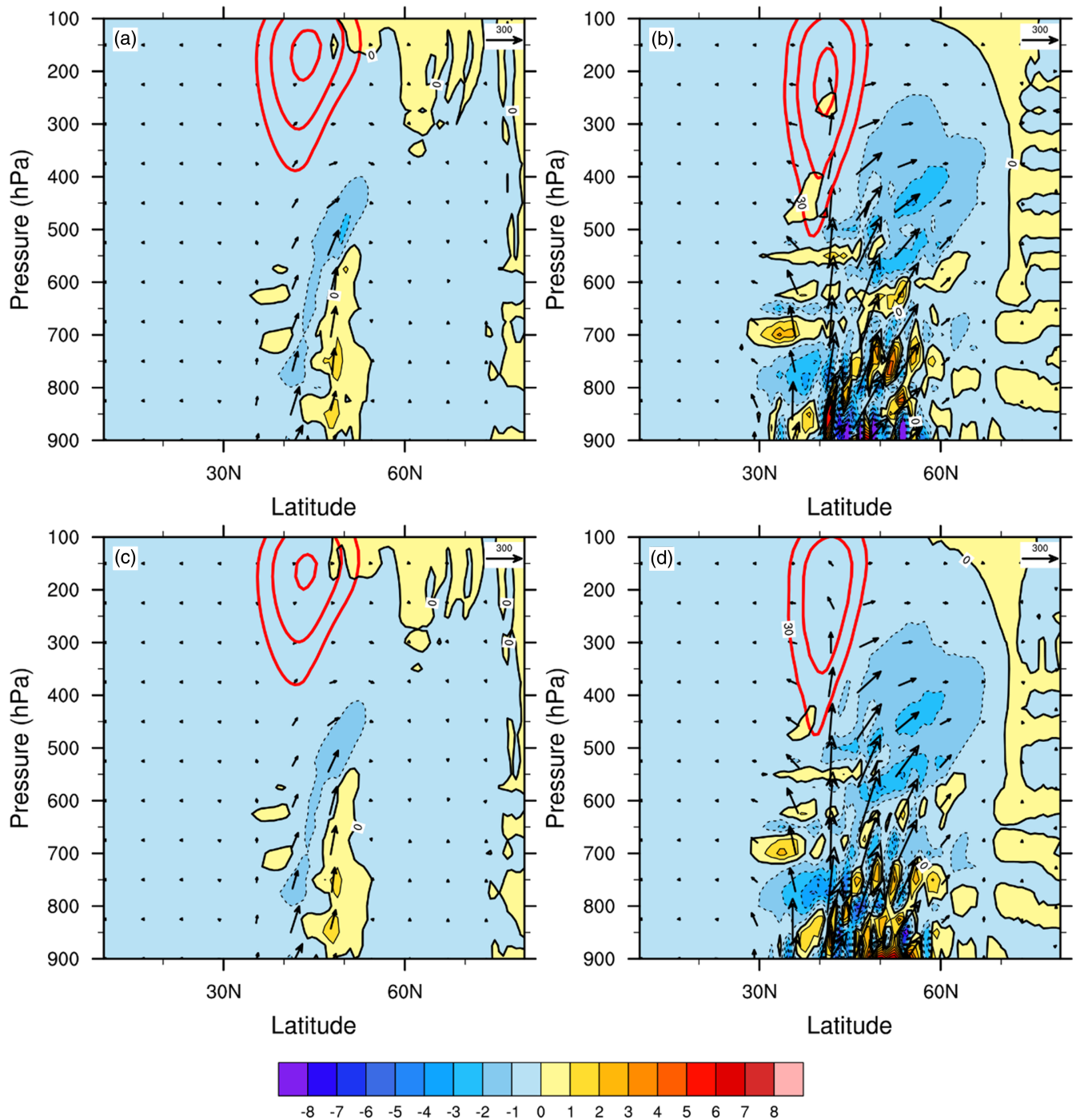


FIGURE 10 Same as Figure 9 but for LC2 [Colour figure can be viewed at wileyonlinelibrary.com]

the EP flux divergence area between 45° and 60° latitudes weaker, thereby leading to a weaker upper-tropospheric jet.

In contrast with LC1, at time $t = 5$ days for LC2 (Figure 10a,c), the signature of upward wave propagation in the lower/middle troposphere is weaker, with flux vectors tending to tilt northward rather than southward. At time $t = 8$ days, compared to LC1, the downward-extended upper tropospheric flux divergence is replaced by some

fragmented areas between 30° and 40° latitudes, which are weaker in the presence of CO_2 . From the latter fragmented areas, the waves are mainly deflected toward the middle to upper tropospheric flux convergence area around 50° and 60° latitudes. Further, associated with the changes in the distribution of the upper tropospheric flux divergence/convergence, the jet shifts southward. Within the upper-tropospheric jet region, there is more southward wave propagation in the experiment with CO_2 . In

general, the impact of CO₂ in weakening the upward wave propagation and the upper-tropospheric jet in LC2 (cf. Figure 10b,d) seems to be stronger than that in LC1.

4 | CONCLUSIONS

In this study, the impacts of varying levels of CO₂ concentration on baroclinic instability development, eddy growth and eddy–mean flow interaction were examined. Considering the slow and gradual changes of forcings together with involvement of various feedbacks in complex systems, such as the atmosphere, makes it hard to disentangle the sole effect of CO₂ on baroclinic eddies. This has provided the motivation to investigate the interaction between radiation and atmospheric flows during the life cycle of baroclinic eddies in experiments with varying levels (0, 250, 500, 750 and 1,000 ppmv) of uniform CO₂ concentrations. The response of the atmosphere to the radiative effects of CO₂ were studied over a period of 16 days by taking into account the different values of the CO₂ concentration in the RRTMG parametrization scheme and carrying out the two main baroclinic life cycles called LC1 and LC2 using the DCASL dynamic core with no involvement of moist and boundary-layer processes. The analytical formulation for initial conditions proposed by Polvani and Esler (2007) was used to set up the LC1 and LC2 experiments.

The results indicate that the radiative heating due to CO₂ is mainly dependent on the temperature structure, which is consistent with the cooling-to-space approximation. The temperature-dependent radiative cooling caused by the presence of CO₂ leads to a decrease in the meridional temperature gradient, thus changing the mean flow and the process of eddy formation. The resulting reduction in the meridional temperature gradient weakens the mean available potential energy, the meridional heat flux and thus the vertical wave propagation, thereby reducing the peak value of EKE and growth rate attained during the life cycle. This summarizes the impact of CO₂ during the stages of surface exponential growth followed by saturation and vertical propagation which is the main contributor to the rapid growth of EKE (Thorncroft *et al.*, 1993). In terms of the peak values of the domain area-average EKE during the 16-day experiments, compared to the experiment with zero CO₂, the relative reductions of 7.4% and 4.2% for, respectively, LC1 and LC2 are observed with the 250 ppmv CO₂ concentration. The additional impact due to increasing values of CO₂ is small but sign-definite; i.e. in the form of reducing EKE (Table 1). Associated with the weakening of vertical propagation of eddies, during the growth stage of EKE, the near-surface EP flux divergence turns to an area of flux convergence under the radiative

heating caused by CO₂. There is also an impact by the eddies on the mean flow, which is more marked in LC2, such that the upper-level jet becomes weaker in the presence of CO₂. While in terms of EKE the radiative impact seems to be stronger in LC1, the wave–mean flow interaction acts in such a way that a greater impact is observed for LC2 during the upper-tropospheric nonlinear saturation stage. Given the dominance of LC1 in the upper troposphere (see Figure 5 of Martius *et al.* (2007)) in the real atmosphere, the greater response can be deduced as being the reduction of EKE. While this may be at least partially related to certain observations like the reduction of eddy heat flux in the Northern Hemisphere during the last four decades (Chemke and Polvani, 2020), caution should be exercised in making arguments for the complex changes to the strength of storm tracks due to the observed climate change (O’Gorman, 2010).

The current study has shed light on the possible radiative impacts of CO₂ during the life cycle of baroclinic eddies in the absence of complex feedback loops that operate in the climate system over longer periods of time, making it hard to disentangle the sole effect of CO₂. Building on the complexity of the model developed to the level of a full general circulation model, we finally stress that the impact of CO₂ on baroclinic instability can be examined in a hierarchy of models to gain understanding of the nature of the radiative forcing of CO₂ in the atmospheric circulation of midlatitudes.

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AUTHOR CONTRIBUTIONS

Mahshid Kaviani: Formal analysis; visualization; writing – original draft. **Farhang Ahmadi-Givi:** Writing – review and editing. **Ali R Mohebalhojeh:** Conceptualization; writing – review and editing. **Daniel Yazgi:** Software.

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REFERENCES

- Adamson, D.S., Belcher, S.E., Hoskins, B.J. and Plant, R.S. (2006) Boundary-layer friction in midlatitude cyclones. *Quarterly Journal of the Royal Meteorological Society*, 132(614), 101–124. <https://doi.org/10.1256/qj.04.145>.

- Andrews, D.G., Holton, J.R. and Leovy, C.B. (1987) *Middle Atmosphere Dynamics*. San Diego, California: Academic Press.
- Barnes, E.A. and Hartmann, D.L. (2011) Rossby wave scales, propagation, and the variability of eddy-driven jets. *Journal of the Atmospheric Sciences*, 68, 2893–2908. <https://doi.org/10.1175/JAS-D-11-039.1>.
- Barnes, E.A. and Polvani, L. (2013) Response of the midlatitude jets and their variability to increased greenhouse gases in the CMIP5 models. *Journal of Climate*, 26, 7117–7135. <https://doi.org/10.1175/JCLI-D-12-00536.1>.
- Ceppi, P. and Shepherd, T.G. (2017) Contribution of climate feedbacks to changes in atmospheric circulation. *Journal of Climate*, 30, 9097–9118. <https://doi.org/10.1175/JCLI-D-17-0189.1>.
- Chang, E.K.M., Guo, Y. and Xia, X. (2012) CMIP5 multimodel ensemble projection of storm track change under global warming. *Journal of Geophysical Research*, 117(D23), D23118. <https://doi.org/10.1029/2012JD018578>.
- Chemke, R. (2017) Atmospheric energy transfer response to global warming. *Quarterly Journal of the Royal Meteorological Society*, 143(706), 2296–2308. <https://doi.org/10.1002/qj.3086>.
- Chemke, R. and Polvani, L.M. (2019) Exploiting the abrupt $4 \times \text{CO}_2$ scenario to elucidate tropical expansion mechanisms. *Journal of Climate*, 32, 859–875. <https://doi.org/10.1175/JCLI-D-18-0330.1>.
- Chemke, R. and Polvani, L.M. (2020) Linking midlatitudes eddy heat flux trends and polar amplification. *npj Climate and Atmospheric Sciences*, 3, 1–8. <https://doi.org/10.1038/s41612-020-0111-7>.
- Clough, S.A., Shephard, M.W., Mlawer, E.J., Delamere, J.S., Iacono, M.J., Cady-Pereira, K., Boukabara, S. and Brown, P.D. (2005) Atmospheric radiative transfer modeling: a summary of the AER codes. *Journal of Quantitative Spectroscopy and Radiative Transfer*, 91, 233–244. <https://doi.org/10.1016/j.jqsrt.2004.05.058>.
- Eckerman, S. (2008) Hybrid σ - p coordinate choices for a global model. *Monthly Weather Review*, 137, 224–245. <https://doi.org/10.1175/2008MWR2537.1>.
- Edmon, H.J., Hoskins, B.J. and McIntyre, M.E. (1980) Eliassen–Palm cross sections for the troposphere. *Journal of the Atmospheric Sciences*, 37, 2600–2616. [https://doi.org/10.1175/1520-0469\(1980\)037<2600:EPCSFT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<2600:EPCSFT>2.0.CO;2).
- Fu, Q. and Liou, K.N. (1992) On the correlated k-distribution method for radiative transfer in nonhomogeneous atmospheres. *Journal of the Atmospheric Sciences*, 49, 2139–2156. [https://doi.org/10.1175/1520-0469\(1992\)049<2139:OTCDMF>2.0.CO;2](https://doi.org/10.1175/1520-0469(1992)049<2139:OTCDMF>2.0.CO;2).
- Hall, N.M.J., Hoskins, B.J., Valdes, P.J. and Senior, C.A. (1994) Storm tracks in a high-resolution GCM with doubled carbon dioxide. *Quarterly Journal of the Royal Meteorological Society*, 120(519), 1209–1230. <https://doi.org/10.1002/qj.49712051905>.
- Harvey, B.J., Shaffrey, L.C. and Woollings, T.J. (2014) Equator-to-pole temperature differences and the extra-tropical storm track responses of the CMIP5 climate models. *Climate Dynamics*, 43, 1171–1182. <https://doi.org/10.1007/s00382-013-1883-9>.
- Hernández-Deckers, D. and von Storch, J.-S. (2010) Energetics responses to increases in greenhouse gas concentration. *Journal of Climate*, 23, 3875–3887.
- Iacono, M.J., Delamere, J.S., Mlawer, E.J., Shephard, M.W., Clough, S.A. and Collins, W.D. (2008) Radiative forcing by long-lived greenhouse gases: calculations with the AER radiative transfer models. *Journal of Geophysical Research*, 113(D13), D13103. <https://doi.org/10.1029/2008JD009944>.
- Jeevanjee, N. and Fueglistaler, S. (2020a) On the cooling-to-space approximation. *Journal of the Atmospheric Sciences*, 77, 465–478. <https://doi.org/10.1175/JAS-D-18-0352.1>.
- Jeevanjee, N. and Fueglistaler, S. (2020b) Simple spectral models for atmospheric radiative cooling. *Journal of the Atmospheric Sciences*, 77, 479–497. <https://doi.org/10.1175/JAS-D-18-0347.1>.
- Kushner, P.J., Held, I. and Delworth, T. (2001) Southern Hemisphere atmospheric circulation response to global warming. *Journal of Climate*, 14, 2238–2249. [https://doi.org/10.1175/1520-0442\(2001\)014<0001:SHACRT>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<0001:SHACRT>2.0.CO;2).
- Lacis, A.A. and Oinas, V. (1991) A description of the correlated k distribution method for modeling nongray gaseous absorption, thermal emission, and multiple scattering in vertically inhomogeneous atmospheres. *Journal of Geophysical Research*, 96(D5), 9027–9063. <https://doi.org/10.1029/90JD01945>.
- Laghaeizadeh, R., Joghataei, M., Yazgi, D. and Mohebalhojeh, A.R. (2020) The diabatic contour-advective semi-Lagrangian algorithms for the dynamical core of global models in terrain-following isentropic and pressure vertical coordinates. *Quarterly Journal of the Royal Meteorological Society*, 146(733), 3715–3728. <https://doi.org/10.1002/qj.3868>.
- Lee, S. and Yoo, C. (2014) On the causal relationship between poleward heat flux and the equator-to-pole temperature gradient: a cautionary tale. *Journal of Climate*, 27, 6519–6525. <https://doi.org/10.1175/JCLI-D-14-00236.1>.
- Lorenz, D.J. and DeWeaver, E.T. (2007) Tropopause height and zonal wind response to global warming in the IPCC scenario integrations. *Journal of Geophysical Research*, 112(D10), D10119. <https://doi.org/10.1029/2006JD008087>.
- Martius, O., Schwierz, C. and Davies, H. (2007) Breaking waves at the tropopause in the wintertime Northern Hemisphere: climatological analyses of the orientation and the theoretical LC1/2 classification. *Journal of the Atmospheric Sciences*, 64, 2576–2592.
- Mlawer, E.J., Taubman, S.J., Brown, P.D., Iacono, M.J. and Clough, S.A. (1997) Radiative transfer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the longwave. *Journal of Geophysical Research*, 102(16), 16663–16682. <https://doi.org/10.1029/97JD00237>.
- Mohebalhojeh, A.R., Joghataei, M. and Dritschel, D. (2016) Toward a PV-based algorithm for the dynamical core of hydrostatic global models. *Monthly Weather Review*, 144, 2481–2502. <https://doi.org/10.1175/MWR-D-15-0379.1>.
- O’Gorman, P.A. (2010) Understanding the varied response of the extratropical storm tracks to climate change. *Proceedings of the National Academy of Sciences of the USA*, 107, 19176–19180.
- O’Gorman, P.A. and Schneider, T. (2008) Energy of midlatitude transient eddies in idealized simulations of changed climates. *Journal of Climate*, 21, 5797–5806.
- Park, S.H., Skamarock, W.C., Klemp, J.B., Fowler, L.D. and Duda, M.G. (2013) Evaluation of global atmospheric solvers using extensions of the Jablonowski and Williamson baroclinic wave test case. *Monthly Weather Review*, 141, 3116–3129. <https://doi.org/10.1175/MWR-D-12-00096.1>.
- Petty, G.W. (2006) *A First Course in Atmospheric Radiation*. Madison, Wisconsin: Sundog Publishing.
- Polvani, L.M. and Esler, J.G. (2007) Transport and mixing of chemical air masses in idealized baroclinic life cycles. *Journal of Geophysical Research*, 112(D23), D23102. <https://doi.org/10.1029/2007JD008555>.
- Stevens, B., Giorgetta, M., Esch, M., Mauritsen, M., Crueger, T., Rast, S., Salzmann, M., Schmidt, H., Bader, J., Block, K., Brokopf,

- R., Fast, I., Kinne, S., Kornbluh, L., Lohmann, U., Pincus, R., Reicheler, T. and Roeckner, E. (2013) Atmospheric component of the MPI-M Earth system model: ECHAM6. *Journal of Advances in Modeling Earth Systems*, 5, 146–172. <https://doi.org/10.1002/jame.20015>.
- Thompson, D.W.J., Wallace, J.M. and Hegerl, G.C. (2000) Annular modes in the extratropical circulation. Part II: trends. *Journal of Climate*, 13, 1018–1036. [https://doi.org/10.1175/1520-0442\(2000\)013<1018:AMITEC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<1018:AMITEC>2.0.CO;2).
- Thorncroft, C.D., Hoskins, B.J. and McIntyre, M.E. (1993) Two paradigms of baroclinic-wave life-cycle behaviour. *Quarterly Journal of the Royal Meteorological Society*, 119(509), 17–55. <https://doi.org/10.1002/qj.49711950903>.
- Thuburn, J. and Craig, G.C. (2002) On the temperature structure of the tropical stratosphere. *Journal of Geophysical Research*, 107(D2), 4017. <https://doi.org/10.1029/2001JD000448>.
- Vallis, G.K. (2017) *Atmospheric and Oceanic Fluid Dynamics. Fundamentals and Large-Scale Circulation*, 2nd edition. Cambridge, UK: Cambridge University Press.
- Vallis, G.K., Colyer, G., Geen, R., Gerber, E., Jucker, M., Maher, P., Paterson, A., Pietschnig, M., Penn, J. and Thomson, S.I. (2018) Isca, v1.0: a framework for the global modelling of the atmospheres of Earth and other planets at varying levels of complexity. *Geoscientific Model Development*, 11, 843–859. <https://doi.org/10.5194/gmd-11-843-2018>.
- Wu, Y., Seager, R., Shaw, T.A., Ting, M.F. and Naik, N. (2013) Atmospheric circulation response to an instantaneous doubling of carbon dioxide. Part II: atmospheric transient adjustment and its dynamics. *Journal of Climate*, 26, 918–935. <https://doi.org/10.1175/JCLI-D-12-00104.1>.
- Yin, J.H. (2005) A consistent poleward shift of the storm track in simulations of 21st century climate. *Geophysical Research Letters*, 32, L18701. <https://doi.org/10.1029/2005GL023684>.
- Yuval, J. and Kaspi, Y. (2016) Eddy activity sensitivity to changes in the vertical structure of baroclinicity. *Journal of the Atmospheric Sciences*, 73, 1709–1726. <https://doi.org/10.1175/JAS-D-15-0128.1>.
- Zappa, G. and Shepherd, T.G. (2017) Storylines of atmospheric circulation change for European regional climate impact assessment. *Journal of Climate*, 30, 6561–6577. <https://doi.org/10.1175/JCLI-D-160807.1>.

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APPENDIX A: FURTHER DETAILS ON INITIALISATION

A complete documentation of the procedure to determine the initial zonally symmetric conditions are presented in Polvani and Esler (2007). For completeness, only an outline is given here. The temperature field is obtained by integrating the thermal-wind equation $\partial T / \partial \varphi = -(H/R)(af + 2u \tan \varphi) \partial u / \partial z$, R being the gas constant for dry air, and considering a reference height-dependent profile such that

$$T(\varphi, z) = T_0 + \frac{\Gamma_0}{(z_T^{-\alpha} + z^{-\alpha})^{1/\alpha}} - \frac{H}{R} \int_0^\varphi (af + 2u \tan \varphi') \frac{\partial u}{\partial z} d\varphi', \quad (\text{A1})$$

where $T_0 = 300$ K, $\Gamma_0 = -6.5 \text{ K} \cdot \text{km}^{-1}$ and $\alpha = 10$. To determine the surface pressure p_s for LC2, the nonlinear equation obtained by setting the surface geopotential to zero,

$$\Phi(\varphi, p_s(\varphi)) = 0, \quad (\text{A2})$$

is solved simply by Taylor expanding the expression for Φ about $p_s = p_0$ leading to the solution $p = p_0 (1 + \Phi_b(\varphi, p_0) / RT(p_0))$ where $T(p_0)$ is the temperature obtained by setting $p_s = p_0$ and Φ_b is the geopotential balanced with the zonal wind u_2 of LC2, that is

$$\Phi_b(\varphi, p) = - \int_0^\varphi (af + 2u_2 \tan \varphi') u_2 d\varphi'.$$

APPENDIX B: SCALING OF FLUX COMPONENTS

Based on Edmon *et al.* (1980), the EP flux on spherical geometry is given by

$$(F^{(\varphi)}, F^{(p)}) = \left(-a \cos \varphi \overline{u'v'}, fa \cos \varphi \frac{\overline{v'\theta'}}{\partial \theta / \partial p} \right), \quad (\text{B1})$$

where a is the radius of the Earth. For a graphical display in latitude–pressure coordinates the flux components are scaled to $(F^{(\varphi)}, F^{(p)})_1$ by

$$(F^{(\varphi)}, F^{(p)})_1 = \cos \varphi \left(\frac{F^{(\varphi)}}{as_\varphi}, \frac{F^{(p)}}{s_p} \right), \quad (\text{B2})$$

where the scale factors $s_\varphi = \pi$ radians and $s_p = 10^5$ Pa are linked to the aspect ratio of the graphic on the page.