



Downstream development of extratropical transition of tropical cyclones over the Southern Hemisphere

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Abstract. Tropical cyclones (TCs) occasionally move into the midlatitudes and transition into extratropical cyclones. The process is known as extratropical transition (ET), which can affect the weather further downstream. The present study conducts a climatological analysis of the dynamics of the downstream development associated with ET over the Southern Hemisphere. These ET events are objectively identified and further grouped into four distinct ET clusters using *K*-means clustering. ET events in clusters 2 and 3 cause pronounced downstream development. The dynamics of the TC-midlatitude flow interaction are diagnosed from both potential vorticity (PV) and eddy kinetic energy (K_e) perspectives. From the PV viewpoint, the anti-cyclonic PV tendency due to the PV advection by diabatically-driven divergent outflow contributes substantially to downstream ridge amplification, supplemented by the nonlinearities arising from PV anomalies advected by flow induced by PV anomalies. From the K_e viewpoint, the transitioning TC injects additional K_e through baroclinic conversion into the midlatitude flow. The energy is then distributed by ageostrophic geopotential fluxes downstream, thereby amplifying the downstream wave packet. We identified two different pathways of downstream development. One pathway (cluster 2) exhibits a classic picture of downstream development, with Rossby wave packets amplifying and propagating farther downstream along a strong and straight jet, whereas the other pathway (cluster 3) is more localised and characterised by a pronounced ridge immediately downstream. These distinct pathways appear to be predominantly determined by the characteristics of the jet, in which the more pronounced downstream ridge is associated with a weak, short, and poleward-orientated jet. Furthermore, our results show that the strong ET events in the southwest Indian Ocean downstream can cause significantly enhanced high-impact weather frequency over Australia.



1 Introduction

20 Tropical cyclones (TCs) occasionally move to the midlatitudes, changing from a warm-core, symmetric structure to a cold-core, asymmetric frontal structure. This process is known as extratropical transition (ET; Klein et al., 2000; Jones et al., 2003; Evans et al., 2017). The transitioning TC can not only cause high-impact weather locally, but can also affect weather downstream from the cyclone. During ET, the transitioning TC impinges on the midlatitude flow, directly modifying the upper-level jet stream and exciting the ridge-trough couplet downstream of the ET. These robust features have been demonstrated from different perspectives in previous studies, including climatological (e.g., Archambault et al., 2013, 2015; Quinting and Jones, 2016), idealised simulations (e.g., Riemer et al., 2008; Riemer and Jones, 2010), and case studies (e.g., Agusti-panareda et al., 2004; Harr and Dea, 2009; Griffin and Bosart, 2014). The upstream ET can lead to subsequent explosive extratropical cyclogenesis in the downstream region (e.g., Agusti-panareda et al., 2004; Cordeira and Bosart, 2010). Downstream heavy precipitation events are additionally associated with the formation of upper-level potential vorticity (PV) streamers, which result from the amplification of a Rossby wave packet through the interaction with an upstream ET (e.g., Grams et al., 2011; Grams and Blumer, 2015). Furthermore, ET events can result in increased forecast uncertainty downstream, with ensemble standard deviation increasing and propagating downstream away from the ET (e.g., Keller et al., 2014; Brannan and Chagnon, 2021). In addition to ET, Parker et al. (2013) have also shown that the outflow from a TC northwest of Australia can contribute to the amplification of Rossby waves and subsequent wave breaking in the southeast of the country, thereby promoting an upper-level anticyclonic PV anomaly, and consequently heatwave development in the region.

There are two main conceptual frameworks used to understand ET and its associated downstream development. The eddy kinetic energy (K_e) framework (Orlanski and Sheldon, 1995) interprets downstream baroclinic wave development as a repeated cycle, in which K_e is dispersed downstream from an existing K_e centre by ageostrophic geopotential fluxes, forming a new energy centre that is further amplified by baroclinic conversion. The new and amplified energy centre subsequently radiates farther downstream and initiates another energy centre. From this viewpoint, TCs can act to provide additional energy, the jet streak is represented as a local maximum of K_e , and the ridge-trough couplet is associated with enhanced K_e during ET (Keller et al., 2019). A very early study by Palmén (1958) found that a large amount of K_e was exported from the region around TC Hazel (1954) to the midlatitude flow during ET. A study of TC David showed that the interaction between the TC and a midlatitude trough to the northwest of the TC resulted in a conversion of available potential energy to kinetic energy, through cold air sinking west of the cyclone and warm air rising to its east (Harr and Elsberry, 2000). More recent studies have established a consistent picture, in which warm air ascending to the east of the transitioning TC produces K_e through baroclinic conversion, while ageostrophic geopotential fluxes redistribute this energy downstream, leading to K_e accumulation on the western flank of the ridge (e.g., Harr and Dea, 2009; Keller et al., 2014; Quinting et al., 2014).

Complementary to the K_e framework, many studies have also examined ET from a PV framework. In this viewpoint, the advective PV tendency is carefully decomposed into that due to different features (PV anomalies) associated with the ET system, namely the cyclonic PV tower that represents the TC, anticyclonic TC outflow, and the diabatically-driven divergent outflow from the TC (e.g., Riemer et al., 2008). This is often achieved by separating the total flow into divergent and non-



divergent (balanced) components using a Helmholtz decomposition, and then by partitioning the balanced flow into those associated with individual PV anomalies using piecewise PV inversion (Davis and Emanuel, 1991). From this perspective, the jet streak formation is interpreted as upper-level frontogenesis, in which the TC divergent outflow enhances the horizontal PV gradient associated with the jet, thereby locally accelerating it (e.g., Riemer and Jones, 2010; Archambault et al., 2013, 2015). The divergent outflow is also important for the ridge-trough couplet developing immediately downstream of ET. Riemer and Jones (2010) conducted an idealised simulation of ET, showing that the divergent outflow made the most dominant contribution to the ridge development during the early stage of the simulation, while the contribution from the TC cyclonic PV tower increased at later stages. From a climatological perspective, Archambault et al. (2013, 2015) showed that the downstream flow amplification over North America is associated with recurving TCs over the western North Pacific that have strong TC–midlatitude flow interactions (measured by the cyclonic PV advection by the divergent outflow). They further found that the divergent flow anchors the downstream ridge, thereby strengthening the pre-existing Rossby wave packets. Similarly, another climatological investigation also documented the importance of cyclonic PV advection by divergent flow to ridge amplification and jet intensification on the western flank of the downstream ridge for the recurving TCs over the western North Pacific that cause Rossby wave amplification downstream (Quinting and Jones, 2016). However, it should be noted that recurving TCs are defined based on changes in their direction of motion (e.g., Burroughs and Brand, 1973), whereas ET events are often defined according to changes in cyclone structure (e.g., Evans and Hart, 2003).

The importance of ET on downstream development has been well documented in the Northern Hemisphere from different perspectives, however, such investigation is very limited in the Southern Hemisphere. Quinting and Jones (2016) conducted a climatology study of the frequency and amplitude of Rossby wave packets downstream of recurving TCs in the south Indian Ocean (30–90° E). They reported a 12–15% increase in Rossby wave packet frequency downstream, with significant increased frequency immediately downstream 2–4 days after recurvature, a relatively weak frequency over Australia, followed by another significant increase over the South Pacific 5–7 days after (as shown in their Fig. 5). Moreover, the amplitude increases significantly downstream over the Australian sector (100–160° E) 3–5 days after the recurvature. A detailed case study of TC Edisoana (1990) in the southwest Indian Ocean was conducted by Griffin and Bosart (2014), showing that a subtropical ridge amplifies downstream of TC Edisoana, associated with the divergent outflow advecting PV poleward, as well as the upper-level PV modification by latent heating. In this event, the anticyclonic PV advection by the divergent wind also acts to strengthen the downstream jet, which then provides baroclinic forcing for the rapid intensification of post-transition Edisoana.

Although Griffin and Bosart (2014) and Quinting and Jones (2016) provide important insights into the downstream development associated with ET in the Southern Hemisphere, there has not yet been a systematic climatological study of the dynamics of ET downstream development over the Southern Hemisphere. Motivated by this gap, the current study provides a climatological study of the dynamics of downstream development associated with ET events in the Southern Hemisphere. The ET events analysed here are taken from a recent study by Jin et al. (2026), which are objectively identified from ERA5 using a cyclone tracking and phase-classification framework (Ullrich et al., 2021; Han and Ullrich, 2025), and subsequently grouped into four distinct clusters based on *K*-means clustering. The paper is organised as follows. Section 2 introduces the datasets and methodologies. Section 3 first examines the seasonality and downstream development of the four ET clusters, then presents



detailed analyses of two strong ET clusters, followed by an influence-focused analysis of strong ET events over the southwest Indian Ocean. Section 4 provides discussions, and Section 5 summarises the results and provides an overview of future work.

90 2 Data and methods

2.1 Data

Daily precipitation and maximum surface temperature are sourced from the Australian Gridded Climate Data (AGCD), with a horizontal grid spacing of 0.05° (Jones et al., 2009). Daily precipitation is accumulated in the period from 0900 (local time) the day before to 0900 (local time) on the current day. The dataset covers the period from 1979 to 2021. To examine the
95 downstream influence of ET on the Australian region, heavy rainfall days and warm days are defined over the land. First, a wet day is defined as a day with daily precipitation of at least 1 mm. At each grid point, a heavy rainfall day is defined as a day on which daily precipitation exceeds the 90th percentile of wet days for the time of year during the 1979-2021 period. Likewise, a warm day is defined as a day with the maximum temperature exceeding the 90th percentile.

The European Centre for Medium-Range Weather Forecasts (ECMWF) v5 (ERA5; Hersbach et al., 2020) is used in this
100 study, with a 6-hourly temporal resolution (00, 06, 12, and 18 UTC) and a horizontal grid spacing of 0.5° . Vertical isobaric levels are considered from 1000 to 100 hPa at 50-hPa intervals, and vertical isentropic levels include 330 and 350 K. These data cover the period from 1979 to 2021.

2.2 Quasigeostrophic omega diagnostic

In adiabatic flow with a sufficiently small Rossby wave number, the quasigeostrophic (QG) omega equation diagnoses the
105 vertical motion from the geopotential fields. Following Hoskins et al. (1978), the $\mathbf{Q}$ -vector form of the QG omega equation is expressed as

$$(\sigma \nabla_p^2 + f_0^2 \frac{\partial^2}{\partial p^2}) \omega = -2 \nabla_p \cdot \mathbf{Q}, \quad (1)$$

where ∇_p represents the horizontal gradient operator on pressure levels, f_0 is the local Coriolis parameter, σ is the static stability, and ω is the diagnosed vertical velocity. $\mathbf{Q}$ is the $\mathbf{Q}$ vector, defined as $\mathbf{Q} = \left[\frac{\partial \mathbf{V}_g}{\partial x} \cdot \nabla \left(\frac{\partial \phi}{\partial p} \right), \frac{\partial \mathbf{V}_g}{\partial y} \cdot \nabla \left(\frac{\partial \phi}{\partial p} \right) \right]$, where $\mathbf{V}_g$
110 is the geostrophic wind, and ϕ is geopotential. Eq. (1) implies that the convergence (divergence) of the $\mathbf{Q}$ vector forces ascent (descent).

The QG vertical motion at any level can be partitioned into that forced by upper-level flow (above 550 hPa) and that forced by lower-level features (below 550 hPa), and therefore, the vertical motion can be partitioned into that forced by upper-level weather feature (e.g., upstream trough), and into that forced by lower-level feature (e.g., front). In the remainder of this study,
115 the QG vertical motion forced by upper levels is termed $\text{QG}\omega_{\text{top}}$, and that forced by lower levels is termed $\text{QG}\omega_{\text{bot}}$. Furthermore, previous studies have shown that the QG diagnostic largely underestimates the vertical motion in regions with strong latent heating (e.g., Quinting and Jones, 2016). Therefore, by comparing the QG omega equation with the total vertical motion (ω),



the importance of the adiabatically-driven vertical motion relative to the vertical motion forced by diabatic processes due to the TC and its interaction with the midlatitude baroclinic zone is evaluated.

120 2.3 PV budget

For the frictionless and adiabatic flow, PV is conserved on isentropic surfaces. The total rate of change of PV is given by

$$\frac{D_\theta}{Dt}q = \frac{\partial}{\partial t}q + \mathbf{V} \cdot \nabla_\theta q = 0, \quad (2)$$

where q is the Ertel PV, $\mathbf{V}$ is the horizontal wind on isentropic surfaces, and ∇_θ denotes the horizontal gradient operator on isentropic levels. The horizontal wind $\mathbf{V}$ can be decomposed into divergent (subscript χ) and non-divergent (subscript ϕ) parts

125 using Helmholtz decomposition as $\mathbf{V} = \mathbf{V}_\chi + \mathbf{V}_\phi$. Therefore, Eq. (2) can be written as

$$\frac{\partial}{\partial t}q + \mathbf{V}_\phi \cdot \nabla_\theta q = -\mathbf{V}_\chi \cdot \nabla_\theta q = I_\chi. \quad (3)$$

The right-hand side term $I_\chi = -\mathbf{V}_\chi \cdot \nabla_\theta q$ is interpreted by Archambault et al. (2013, 2015) and O'Brien and Reeder (2017) as the interaction of the divergent outflow from a TC with the midlatitude waveguide.

From Barnes et al. (2025), the PV budget can be obtained by linearising Eq. (3) by partitioning the fields as the sum of a
130 time-zonal mean and a perturbation, e.g., $q = \bar{q} + q'$, and $\mathbf{V}_\phi = \overline{\mathbf{V}_\phi} + \mathbf{V}_\phi'$. The time mean is defined here as a 30-day running mean centred on the analysis time. Equation (2) can then be expressed as

$$\begin{aligned} \frac{\partial q'}{\partial t} + \overline{\mathbf{V}_\phi} \cdot \nabla_\theta q' + \mathbf{V}_\phi' \cdot \frac{\partial \bar{q}}{\partial y} &= -\mathbf{V}_\chi \cdot \nabla_\theta q - \mathbf{V}_\phi' \cdot \nabla_\theta q' \\ &= I_\chi + I_\phi'. \end{aligned} \quad (4)$$

The left-hand side of Eq. (4) describes a barotropic Rossby wave on isentropic levels. On the right-hand side, the nonlinear term $I_\phi' = -\mathbf{V}_\phi' \cdot \nabla_\theta q'$ is interpreted as the advection of PV anomalies by surrounding PV anomalies. In this PV framework, the
135 barotropic Rossby wave is impacted by I_χ and I_ϕ' . I_χ has been related to ET, in which the divergent outflow of a transitioning cyclone advects the PV gradient further poleward, thereby amplifying the Rossby wave packet. Riemer and Jones (2010) have shown that this mechanism contributes substantially to the ridge building downstream of ET.

2.4 K_e budget

In addition to the PV viewpoint, ET can also be explained from a K_e framework. In this framework, midlatitude ridge-trough
140 couplets are characterised by enhanced K_e , a jet streak features a local K_e maximum, and the transitioning cyclone contributes to the amplification of the midlatitude wave by injecting additional energy (e.g., Harr and Dea, 2009; Keller et al., 2019).

The K_e budget is calculated by partitioning the field as the sum of a time-mean (subscript m) and a perturbation. The horizontal wind on isobaric surfaces is defined as $\mathbf{V} = \mathbf{V}_m + \mathbf{v}$, and likewise, scalar variables are defined as $A = A_m + a$. Following Harr and Dea (2009) and Quinting and Jones (2016), the time-mean is defined as a 30-day running mean centred on
145 the analysis time. From Orlanski and Sheldon (1995), the K_e budget can be expressed as

$$\frac{\partial K_e}{\partial t} = -\nabla_p \cdot (\mathbf{v}\phi)_a - \omega\alpha - \nabla_p \cdot (\mathbf{V}K_e) - \mathbf{v} \cdot (\mathbf{v} \cdot \nabla_p \mathbf{V}_m) - \frac{\partial(\omega\phi)}{\partial p} - \frac{\partial(\omega K_e)}{\partial p} + \text{residual}, \quad (5)$$



where $K_e = \frac{1}{2}|\mathbf{v}|^2$, ϕ denotes the geopotential height, ω is the vertical velocity, ∇_p represents the horizontal gradient operator on pressure levels, and α is the specific volume. On the right-hand side, the first term $-\nabla_p \cdot (\mathbf{v}\phi)_a$ describes the ageostrophic (subscript a) geopotential eddy flux divergence. The second term $-\omega\alpha$ is the baroclinic conversion from available potential energy to K_e . The third term $-\nabla_p \cdot (\mathbf{V}K_e)$ is the divergence of the K_e flux, describing the propagation of K_e by the total wind. The fourth term, $-\mathbf{v} \cdot (\mathbf{v} \cdot \nabla_p \mathbf{V}_m)$, is the barotropic conversion term, representing the conversion of K_e between the eddies and the mean flow.

As calculated in previous studies (Harr and Dea, 2009; Quinting and Jones, 2016), Eq. (5) is vertically integrated between 1000 and 100 hPa to take account of the entire troposphere. The integrals of the vertical flux terms (the fifth and sixth terms on the right-hand side of Eq. (5)) are small and are neglected. The integrals are then normalised by the gravitational constant g , which gives the final expression of the K_e budget as

$$\frac{1}{g} \int_{1000}^{100} \frac{\partial K_e}{\partial t} dp = \frac{1}{g} \int_{1000}^{100} \left[-\nabla_p \cdot (\mathbf{v}\phi)_a - \omega\alpha - \nabla_p \cdot (\mathbf{V}K_e) - \mathbf{v} \cdot (\mathbf{v} \cdot \nabla_p \mathbf{V}_m) + \text{residual} \right] dp. \quad (6)$$

The normalisation defines units of $J m^{-2}$ for the vertically-integrated K_e and $W m^{-2}$ for its budget terms on the right-hand side in Eq. (6). For simplicity, we refer to the ageostrophic geopotential eddy flux as *ageoflux*, the second integral on the right-hand side (i.e., baroclinic conversion) as *bcli*, and the fourth integral (i.e., barotropic conversion) as *brtr* in the remainder of this study.

2.5 Statistical significance

To illustrate the downstream influences of ET on the Australian region, the frequency of warm-day and heavy-rainfall-day anomalies (as defined in Section 2.1) are calculated as a deviation from a 30-day running mean centred on the analysis time. The statistical significance of the anomalies is tested via a Monte Carlo analysis. We create one thousand random composites of anomalies. Each composite has a number of events centred on a date randomly selected from a randomly chosen year from the 1979-2021 period and a randomly chosen month and day from a 14-day window around the ET time. We choose the upper and lower 5% percentiles of the Monte Carlo composites, and the anomalies are significant if they are either greater or less than these percentiles.



170 3 Results

3.1 ET clusters

As defined by Jin et al. (2026), transitioning TCs are categorised into four ET clusters. The four clusters exhibit differences in structural changes, background environment, and interaction with the midlatitude flow (see Fig. 11 in Jin et al. (2026)). As shown in Fig. 1, clusters C1 and C4 follow the conventional ET seasonality, with more events occurring at lower latitudes during January and February, and fewer events occur at higher latitudes in the austral autumn. In clusters C2 and C3, however, most ET events occur in the period from December to March, with a more poleward location (Fig. 1b and c). On average, there is no apparent seasonality difference between C2 and C3.

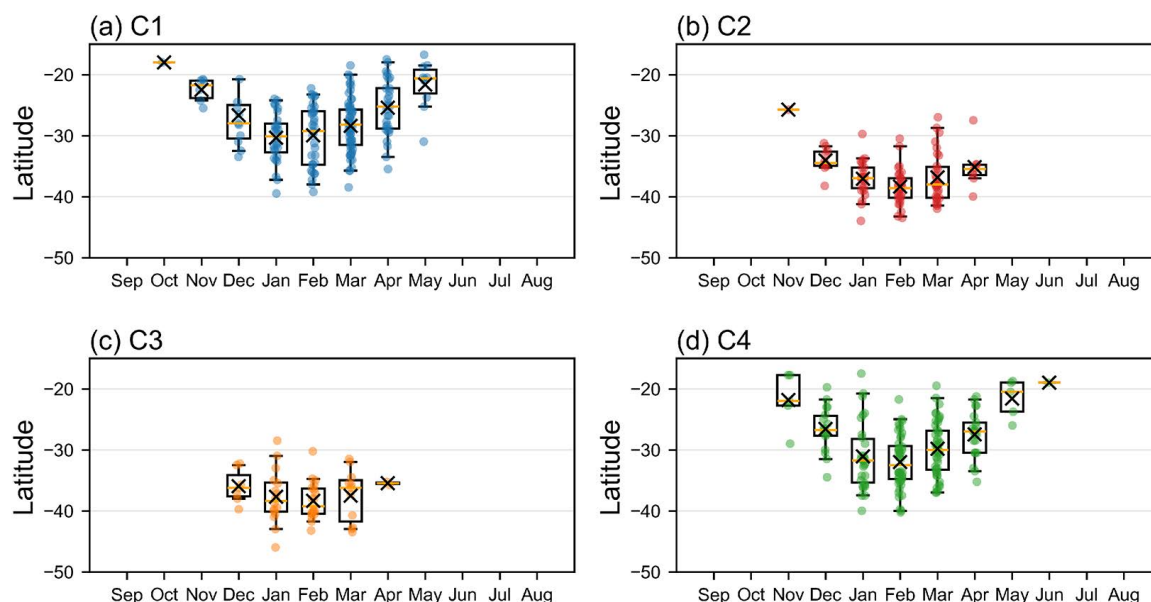


Figure 1. Latitude of cyclones at their ET time (dots) by month and corresponding box and whisker plots for the four ET clusters (a-d). The medians are marked by orange lines, the means by black crosses, the interquartile ranges by boxes, and the whiskers indicate the 1st and 99th percentiles. Both of the monthly statistics are based on the period 1979-2021.



3.2 Downstream development in ET clusters

To examine the downstream development for the four ET clusters, we compute Hovmöller diagrams of 330-K PV and mean sea level pressure (MSLP) anomalies. For each cluster, Hovmöller diagrams are generated for individual events centred on ET longitude and time, latitudinally averaged over a band extending $\pm 15^\circ$ from the ET latitude.

Clusters C2 and C3 show that the transitioning cyclone lies along the leading edge of a large cyclonic PV anomaly and upstream of a strong anticyclonic PV anomaly (Fig. 2b and c). This reflects that the transitioning TC is embedded in highly-amplified midlatitude flow, with a trough approaching from the west and a ridge building downstream. As a comparison, the magnitudes of the upstream MSLP anomaly and downstream PV anomaly are much weaker in C1 and C4 (Fig. 2a and d), suggesting little downstream amplification in these clusters. Therefore, the subsequent analysis focuses on C2 and C3.

In C2, alternating anticyclonic and cyclonic PV and MSLP anomalies extend 60° upstream to 120° downstream of the ET location and persist for about 10 days (Fig. 2b). These anticyclonic and cyclonic PV anomalies constitute an eastward-propagating Rossby wave packet, with an estimated group speed of $45^\circ \text{ day}^{-1}$ based on the leading edge of the wave packet. After ET, a large longitudinal displacement in the TC position is observed (Fig. 2b), with approximately 28° within 48 h. This indicates an increase in TC translation speed after ET, plausibly because the transitioning TC becomes coupled with the midlatitude jet. The first downstream anticyclonic PV anomaly peaks one day after ET, while the first downstream cyclonic PV anomaly remains strong between ET and +24 h and then decays. A negative MSLP anomaly lies east of the downstream cyclonic anomaly, indicating likely surface cyclogenesis ahead of the downstream trough. Far downstream, the second downstream anticyclonic PV anomaly forms to the east of the surface low.

In C3, the magnitude of the first downstream anticyclonic PV anomaly is much larger (Fig. 2c) than that in C2 (Fig. 2b). The anticyclonic PV anomaly exhibits a low phase speed, with about 6° day^{-1} (Fig. 2c). The associated Rossby wave packet also appears to propagate downstream relatively slowly. The first downstream cyclonic PV anomaly is weak, while the negative MSLP anomaly does not occur far downstream until +48 h. Nonetheless, the second downstream anticyclonic PV anomaly appears to be stronger than that in C2 (Fig. 2 b and c). Similar patterns are observed on 350-K surface, with a stronger first downstream anticyclonic PV anomaly in the two clusters (Fig. S1).

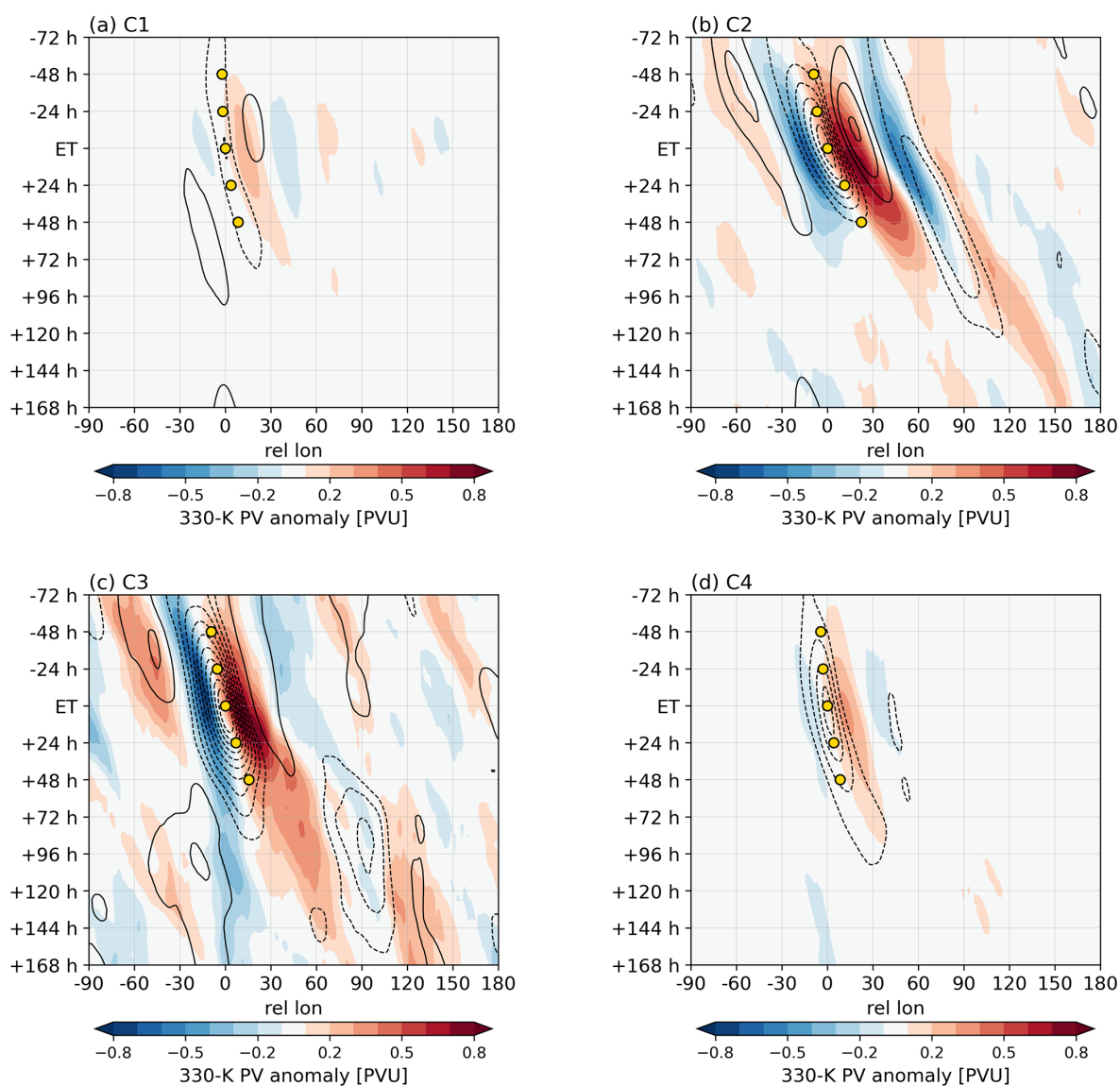


Figure 2. ET-relative composite Hovmöller diagrams of 330-K PV anomaly (shaded as per the colorbar) and MSLP anomaly (black contours every 1 hPa starting from ± 1 hPa) for the four ET clusters (a-d). PV and MSLP anomalies are calculated relative to the 30-day running mean. Yellow dots mark the TC position.



3.3 Downstream development in C2 and C3

3.3.1 Extratropical flow response

For C2 and C3, lead-lag composites of 350-K PV, PV anomaly, and MSLP anomaly are generated relative to ET time and position. Consistent with Fig. 2b, C2 shows an eastward-propagating couplet of lows and highs at the surface and upper levels. From -48 hr to ET (Fig. 3a-c), the cyclone moves southeastward and interacts with the midlatitude flow on its southeastern side. Meanwhile, PV contours overturn anticyclonically downstream, forming a highly amplified PV ridge and an elongated PV streamer on its eastern flank. The downstream cyclonic PV anomaly filament appears to support surface cyclogenesis, indicated by an enhanced negative MSLP anomaly from ET onwards (Fig. 3c). From ET onwards, the Rossby wave packet, marked by cyclonic and anticyclonic PV anomalies, propagates eastward along the waveguide (where the PV gradient is large), reaching around 80° E (Fig. 3c-e).

In C3, the transitioning TC and downstream ridge are much stronger and broader from ET onwards than those in C2 (Fig. 3f-h). Between -24 hr and ET, the downstream PV contours bend poleward dramatically, accompanied by an enhanced anticyclonic PV anomaly (Fig. 3g-h). Unlike C2, C3 does not develop a distinct elongated PV streamer on the eastern flank of the ridge. After ET, C3 shows that the anticyclonic PV anomaly extends poleward and eastward over a large downstream area by +48 hr (Fig. 3h-j). In line with the composite Hovmöller diagrams (Fig. 2c), the downstream anticyclonic PV anomaly is more stationary. At the surface, the wave packet, marked by negative and positive MSLP anomalies, exhibits poleward propagation in C3 (Fig. 3h-j), rather than propagating zonally as in C2 (Fig. 3c-e).

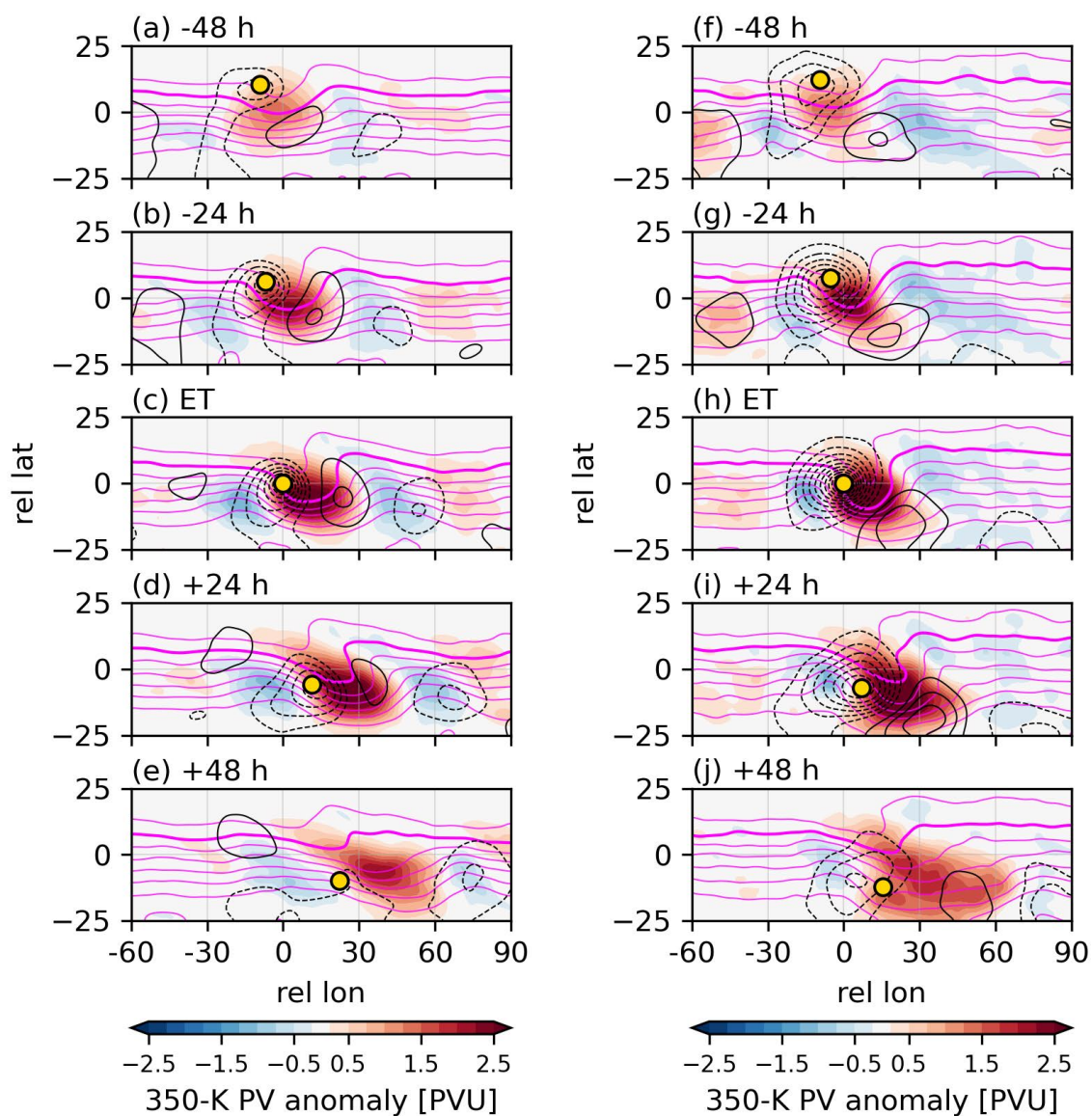


Figure 3. Lead-lag composites of 350-K PV (magenta contours every 1 PVU; $1 \text{ PVU} = 10^{-6} \text{ K m}^2 \text{ kg}^{-1} \text{ s}^{-1}$), PV anomaly (shaded), and MSLP anomaly (black contours every 2 hPa starting from ± 2 hPa; solid for positive and dashed for negative values) for (a-e) C2 and (f-j) C3. PV and MSLP anomalies are calculated relative to the 30-day running mean. Yellow dots mark the TC position.



3.3.2 QG diagnostics

220 The QG omega equation (described in Section 2.2) is used to examine the importance of the adiabatically-driven vertical motion forced by features on different pressure levels. We then compare the QG omega to the total vertical motion (ω) from ERA5 to evaluate the importance of the vertical motion forced by diabatic processes.

In C2, the transitioning TC approaches the equatorward side of the jet entrance, while the jet downstream of the TC intensifies and curves anticyclonically (Fig. 4a-c). A dipole of $QG\omega_{top}$ coincides with the upstream trough to the southwest of the TC at
225 -24 hr, and strengthens and reaches its maximum magnitude at ET (Fig. 4a and b). Another $QG\omega_{top}$ dipole develops near the downstream trough (Fig. 4b and c), with ascent at the leading edge of the trough favouring surface cyclogenesis (Fig. 3c and d). At the lower levels, the TC-related cyclonic circulation interacts with the baroclinic zone, inducing warm advection to the southeast and cold advection to the northwest of the transitioning cyclone, marked by the poleward and equatorward deflection of the temperature contours respectively. The resulting thermal advection forces ascent and descent in the corresponding regions
230 (Fig. 4d-f). At ET, the $QG\omega_{bot}$ dipole is constrained near the transitioning cyclone (Fig. 4e). The spatial pattern of ω is similar to that associated with QG forcing (Fig. 4g-i), however, the magnitude of ω is around 6-8 times greater than $QG\omega_{bot}$ and $QG\omega_{top}$. This result suggests that a large amount of diabatically-driven vertical motion is linked to the transitioning TC, which is particularly pronounced at ET time. ω associated with the TC appears to be diabatically dominant before and at ET, whereas it becomes adiabatically dominant after ET (Fig. 4). Furthermore, the total vertical motion near the TC is primarily contributed
235 by the lower-level forcing (Fig. 4d-e) and supplemented by the upper-level forcing (Fig. 4a-b) from -24 hr to ET.

C3 features a weaker and wavier downstream jet associated with a deep downstream ridge (Fig. 5a-c), compared with the broad and more zonally-oriented downstream jet in C2 (Fig. 4a-c). The upstream $QG\omega_{top}$ dipole aligning with a highly amplified upstream trough is stronger than in C2, while the downstream $QG\omega_{top}$ dipole develops and peaks after ET (Fig. 5a-c). C3 shares a similar pattern of lower-level temperature and QG forcing with C2, but with a larger amplitude (Fig. 5d-f).
240 At ET, a broad and strong $QG\omega_{bot}$ dipole (Fig. 5e) is consistent with the intense and spatially extended TC (Fig. 3h). Unlike C2, the lower- and upper-level forcing contribute approximately equally to the total vertical motion in C3 (Fig. 5).

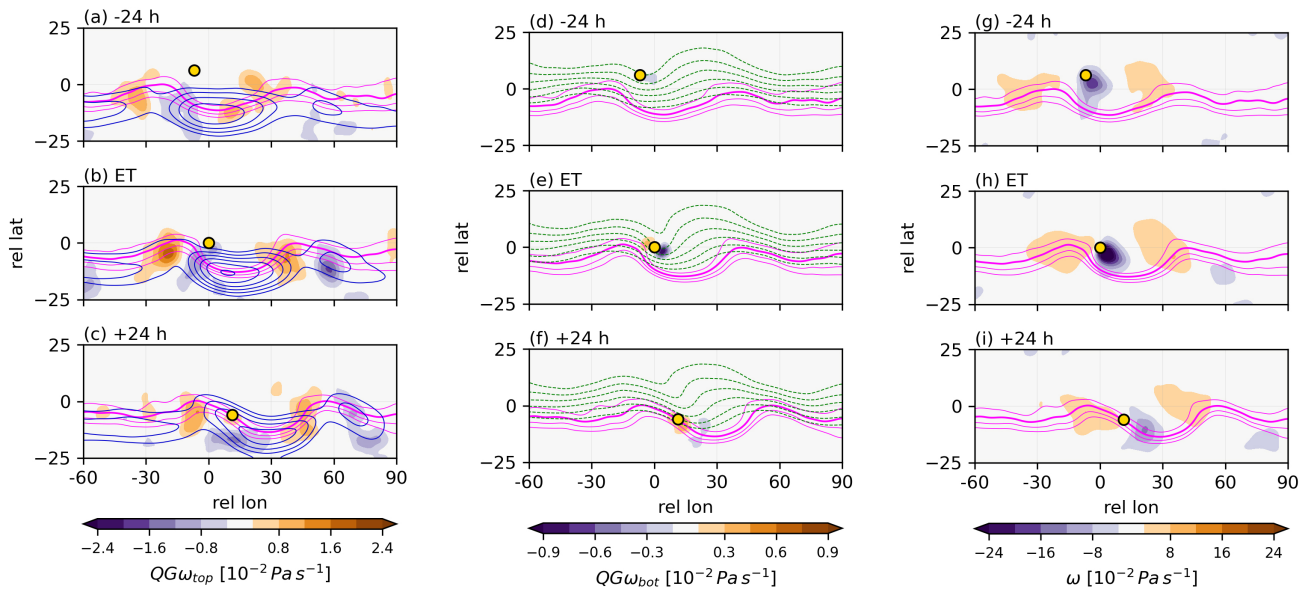


Figure 4. Lead-lag composites for C2 show (a-c) 500-700-hPa averaged $QG\omega_{top}$ (shaded as per the colourbar) and 200-300-hPa averaged wind speed (blue contours every 4 m/s starting from 28 m/s); (d-f) 500-700-hPa averaged $QG\omega_{bot}$ (shaded as per the colourbar) and 800-900-hPa averaged temperature (green dashed contours every 2 K starting from 280 K); (g-i) 500-700-hPa averaged ω (shaded as per the colourbar). The 200-300-hPa averaged PV is indicated by the magenta contours (at -1.5, -2, -2.5, and -3 PVU) in (a-i). Yellow dots mark the TC position.

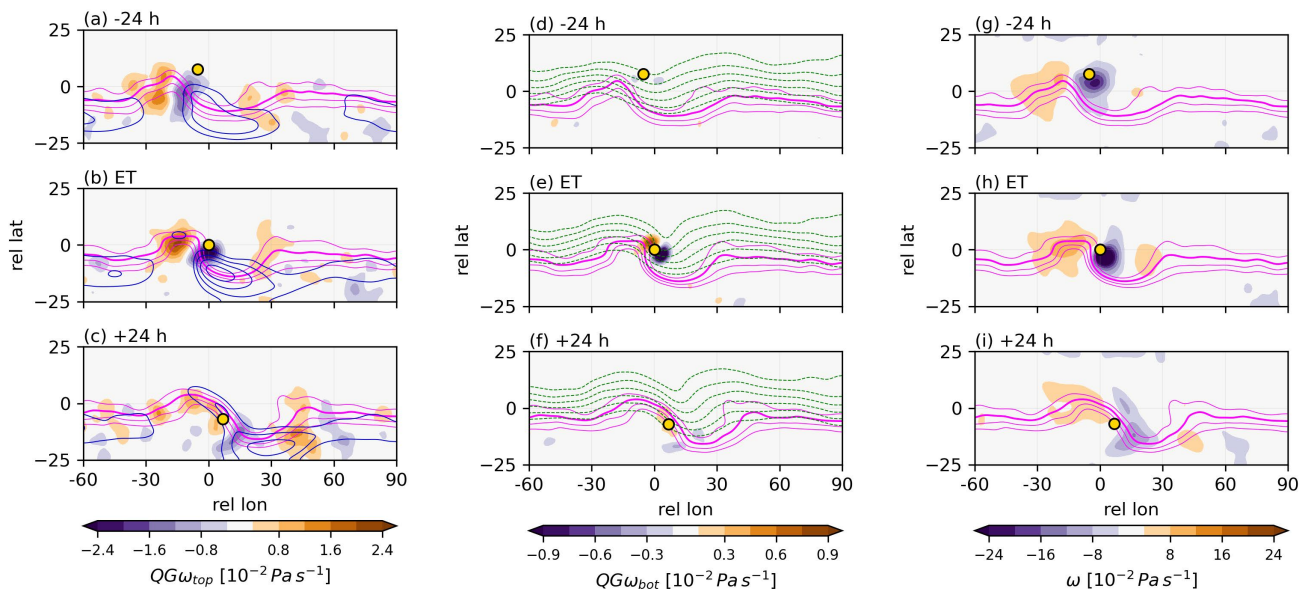


Figure 5. As in Fig. 4, but for C3.



3.3.3 PV budget

To understand the dynamics of downstream development associated with ET, the PV budget (described in Section 2.3) is applied to C2 and C3. The interaction term I_χ , nonlinear term I'_ϕ , and propagation terms $[-\overline{\mathbf{V}_\phi} \cdot \nabla_\theta q' - \mathbf{V}_\phi' \frac{\partial \bar{q}}{\partial y}]$ in Eq. 4 are calculated on the 350-K isentropic surface. Composites of these terms are shown, together with the divergent wind $\mathbf{V}_\chi$, non-divergent wind $\mathbf{V}_\phi$, and PV on the same isentropic surface, for C2 and C3 in Figs. 6 and 7 respectively.

In C2, the transitioning TC impinges on the waveguide, with a PV ridge forming immediately downstream of the TC, and a PV streamer on its eastern flank intruding equatorward and wrapping anticyclonically (Fig. 6). During the 48-hr window centred on ET, the Rossby wave packet propagates downstream along the waveguide, reaching to 90° east of the ET location (Fig. 6k-o). A region of anticyclonic PV tendency associated with I_χ lies on the western flank of the downstream ridge, remaining strong between -24 h and ET and weakening rapidly after ET (Fig. 6a-c). The anticyclonic tendency associated with I_χ is presumably the result of the divergent outflow from convection associated with the transitioning TC, advecting the PV gradient (and therefore the waveguide) southwestward. Like I_χ , the divergent outflow is particularly strong before and during ET and weakens substantially afterwards, presumably the result of reduced convection in the TC after ET (Fig. 6a-c). The westward component of the divergent outflow lies perpendicular to the upstream trough and acts against the eastward propagation of the Rossby wave packet embedded in the non-divergent flow (Fig. 6k-o).

As shown in Fig. 6f-i, the nonlinear term I'_ϕ produces an anticyclonic PV tendency on the southern edge of the PV ridge, likely associated with poleward advection of the PV contours by the cyclonic circulation of the TC PV tower. In contrast, a cyclonic PV tendency occurs around the transitioning TC, presumably reflecting the downstream advection of cyclonic PV by the approaching upstream trough. From -24 to +12 hr, both I_χ and I'_ϕ have signs opposite to the propagation terms $[-\overline{\mathbf{V}_\phi} \cdot \nabla_\theta q' - \mathbf{V}_\phi' \frac{\partial \bar{q}}{\partial y}]$ in the southern and western sides of the downstream ridge. The TC divergent outflow from the TC, together with nonlinear interactions between TC and midlatitude features, counteract the linear propagation-related PV tendencies, while reinforcing the existing anticyclonic PV anomaly. These processes therefore slow the eastward-propagating downstream ridge and promotes its amplification. At +24 hr, however, I'_ϕ produces a cyclonic tendency within the downstream ridge and an anticyclonic tendency within the downstream trough (Fig. 6j). The tendency pattern is broadly in phase with the propagation terms (Fig. 6o). The weakening TC cyclonic PV tower and its associated circulation is likely responsible for the reduced nonlinear PV advection, and therefore, the downstream wave pattern becomes increasingly controlled by the linear Rossby wave propagation.

Figure 7 shows the composite analysis for C3. In contrast to C2, the Rossby wave packet propagates slowly eastward along a weaker background PV gradient (Fig. 7k-o). A major difference from C2 is the absence of pronounced propagation terms between 60 and 90° E, which reflects that the downstream development is more spatially localised in C3. Compared with C2, C3 is featured by a stronger and broader anticyclonic PV tendency produced by I_χ , together with more intense divergent outflow (Fig. 7a-c). These features suggest more intense convection associated with the transitioning TC, particularly before and during ET. Moreover, the divergent outflow has more southward component (perpendicular to the downstream ridge) in



275 C3, thereby favouring a more poleward-extending ridge structure. After ET, unlike C2, the I_χ -produced anticyclonic tendency in C3 extends over a broad downstream region, despite its weaker magnitude.

In C3, the nonlinear term I'_ϕ produces a strong anticyclonic tendency along the southwestern flank of the ridge (Fig. 7f-j). This tendency is stronger and persists longer than its counterpart in C2 throughout the analysis period. As in C2, I_χ and I'_ϕ in C3 substantially slow the eastward propagation of the Rossby wave packet. However, different to C2, the nonlinear PV
280 tendencies in C3 remain out of phase with the propagation terms at +24 hr (Fig. 7j and o), suggesting that nonlinear processes continue to influence the downstream wave. The sustained anticyclonic tendency by I'_ϕ may reflect the persistence of the TC cyclonic PV tower and its associated circulation, which continue to advect the surrounding PV anomalies.

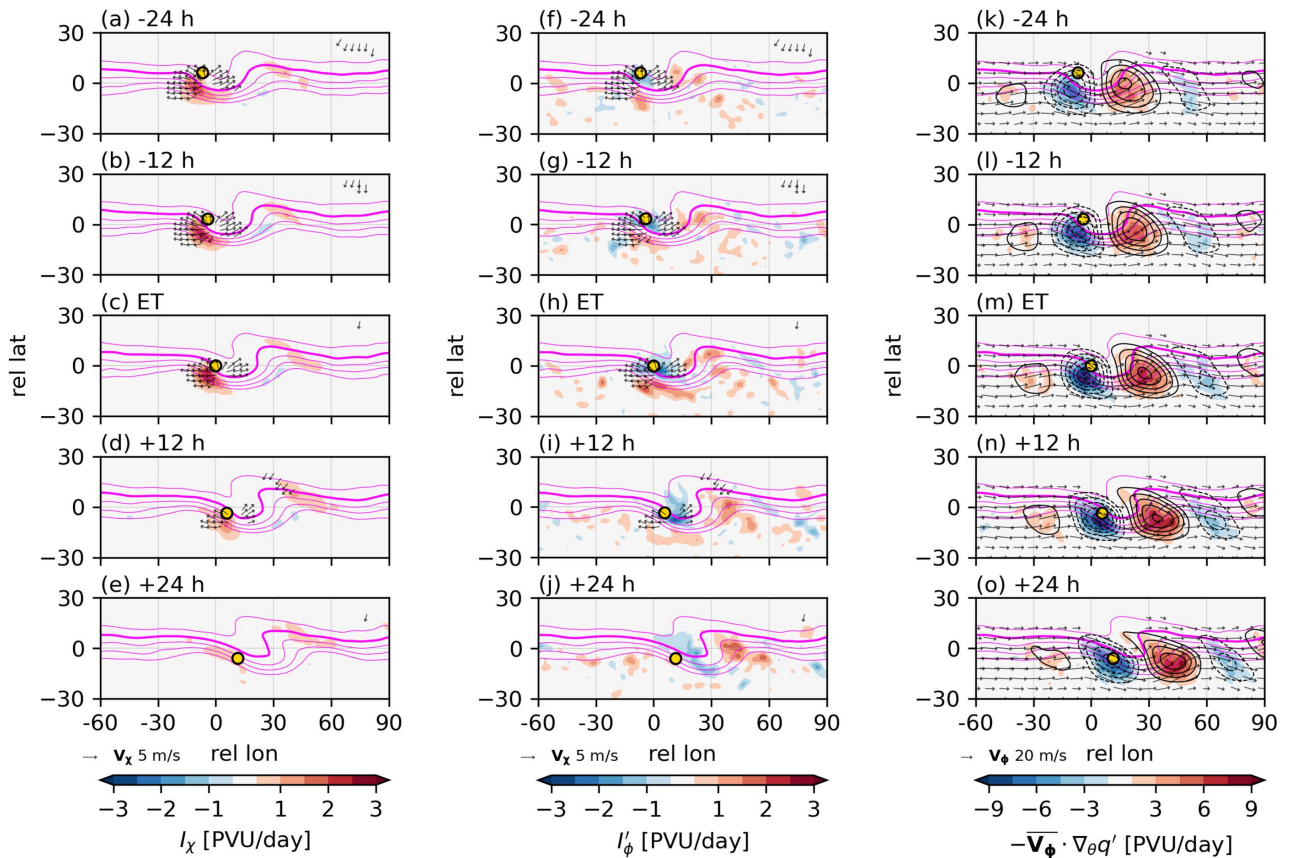


Figure 6. Lead-lag composites of PV budget (Eq. 4) for C2 show (a-e) I_χ (shaded as per the colourbar), (f-j) I'_ϕ (shaded as per the colourbar), and (k-o) $-\mathbf{V}_\phi \cdot \nabla_\theta q'$ (shaded as per the colourbar) and $-\mathbf{V}_\phi' \frac{\partial q}{\partial y}$ (black contours every 1.5 PVU/day starting at ± 1.5 PVU/day). PV is indicated by the magenta contours (at -1, -2, -3, -4 and -5 PVU) in (a-i), with the thick contour indicating -2 PVU. Divergent wind vectors $\mathbf{V}_\chi$ (only for vector > 2.5 m/s) are displayed in (a-j), and nondivergent wind vectors $\mathbf{V}_\phi$ (only for vector > 10 m/s) are displayed in (k-o). All composites are on the 350-K isentropic. Yellow dots mark the cyclone position.

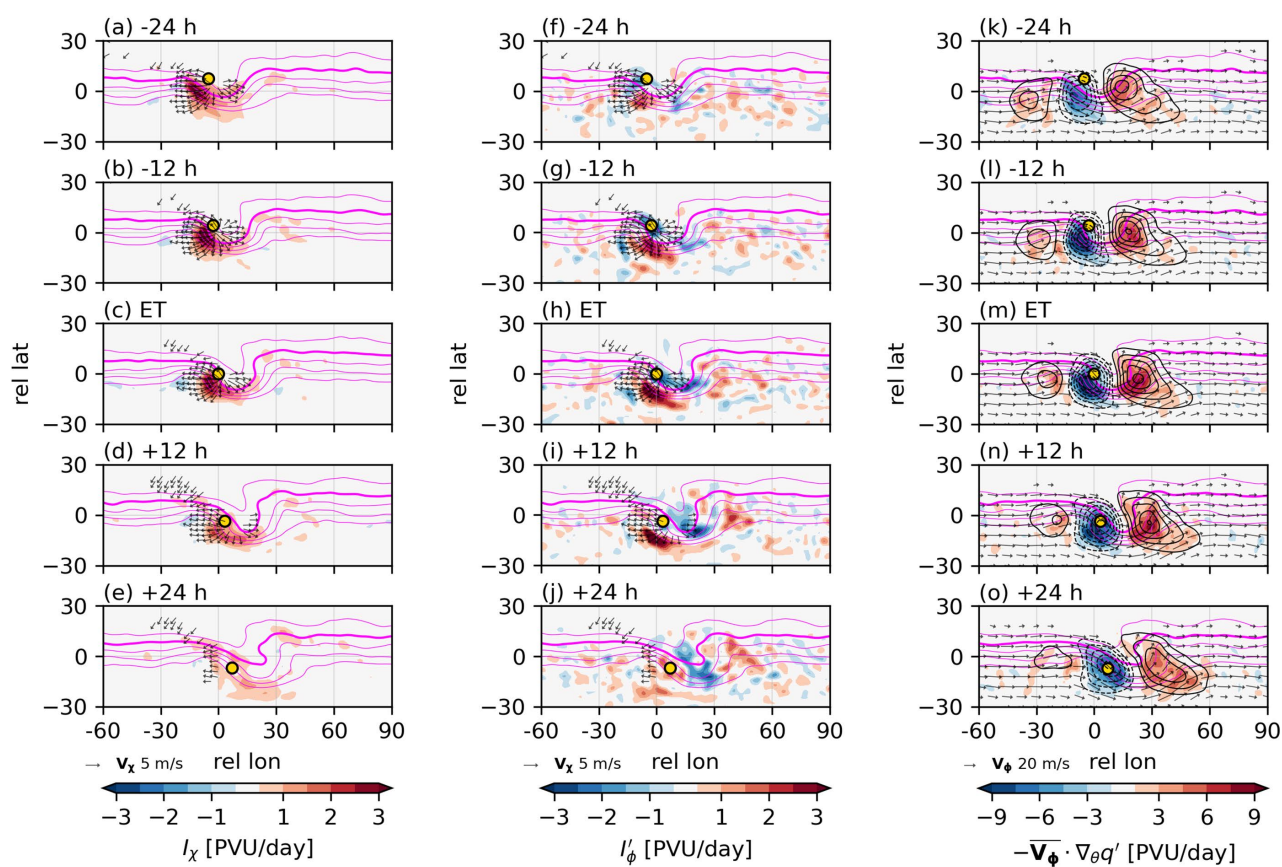


Figure 7. As in Fig. 6, but for C3.



3.3.4 K_e budget

Following previous studies (e.g., Harr and Dea, 2009; Quinting and Jones, 2016), the downstream development of C2 and C3 ET events is also analysed from a K_e framework (described in Section 2.3). For C2 and C3, composites are produced for eddy kinetic energy K_e , baroclinic conversion $bcli$, barotropic conversion $brtr$, ageostrophic geopotential eddy flux $ageoflux$, K_e flux by the total wind, and 500-hPa geopotential height (Z500).

The K_e budget for C2 is shown in Fig. 8. Positive $bcli$ lies to the south-southeast of the TC, strengthening before ET, peaking at ET time and decaying afterwards (Fig. 8a-e). During this process, $bcli$ occurs as a result of rising warm air ahead of the transitioning cyclone (Fig. 4), and it contributes to the immediate growth of K_e and to the formation of a local maximum on the western flank of the downstream ridge (Fig. 8a-e). As shown in Fig. 8f-j, $ageoflux$ divergence coincides with the region of strong $bcli$, suggesting that the additional energy generated by baroclinic conversion is transported downstream in the midlatitude flow. At -24 hr, a weak $ageoflux$ already occurs between 30 and 60° E, far downstream of the transitioning TC (Fig. 8f). The weak flux may be associated with a pre-existing midlatitude baroclinic disturbance (e.g., an extratropical cyclone), which already transfer energy downstream before ET. As the TC undergoes the transition, the $ageoflux$ clearly strengthens in the far downstream region (Fig. 8g-j). Between ET and +24 hr, energy rapidly accumulates on the eastern flank of the downstream ridge and forms a secondary K_e maximum (Fig. 8c-e), due to the convergence of $ageoflux$ (Fig. 8h-j). Meanwhile, a weak K_e centre develops farther downstream at around 70° E, ahead of the downstream trough and colocated with another $ageoflux$ convergence. This highlights the dominant role of the ageostrophic geopotential flux in propagating K_e downstream, consistent with previous studies (e.g., Orlanski and Sheldon, 1995; Harr and Elsberry, 2000; Quinting and Jones, 2016). Furthermore, the convergence and divergence of the K_e flux by the total wind are generally located around the K_e maximum, and the flux vectors suggest the K_e is transported through the base of the ridge and toward the trough farther downstream (Fig. 8k-o). The remaining term $brtr$ has relatively lower magnitude compared with $bcli$ and the convergence of $ageoflux$, thereby contributing little to downstream K_e generation (Fig. S2).

In contrast with C2, the budget terms in C3 have larger magnitudes and are more spatially confined in the downstream area. Between -24 h and ET, positive $bcli$ strengthens and becomes concentrated within a small region on the southeast of the transitioning TC, while K_e intensifies in the region where the TC is captured by a highly-amplified upstream trough (Fig. 9a-c). After ET, both $bcli$ and K_e expand spatially in this coupled TC-trough region. Although $bcli$ weakens, K_e continues to increase (Fig. 9d-e). As shown in Fig. 9f-j, the co-location of strong $ageoflux$ divergence and $bcli$ surrounding the TC suggests that locally generated K_e is rapidly transported primarily to the western side of the downstream ridge. However, the $ageoflux$ and its convergence in C3 are much weaker east of 30° E than those in C2, indicating the limited role of ageostrophic geopotential flux in propagating K_e far downstream. The convergence and divergence of the K_e flux in C3 predominantly lie over the upstream trough and downstream ridge, but they are relatively weak at 50° E (Fig. 9k-o). As in C2, the $brtr$ contributes little to downstream K_e generation compared with other terms in C3 (Fig. S3).

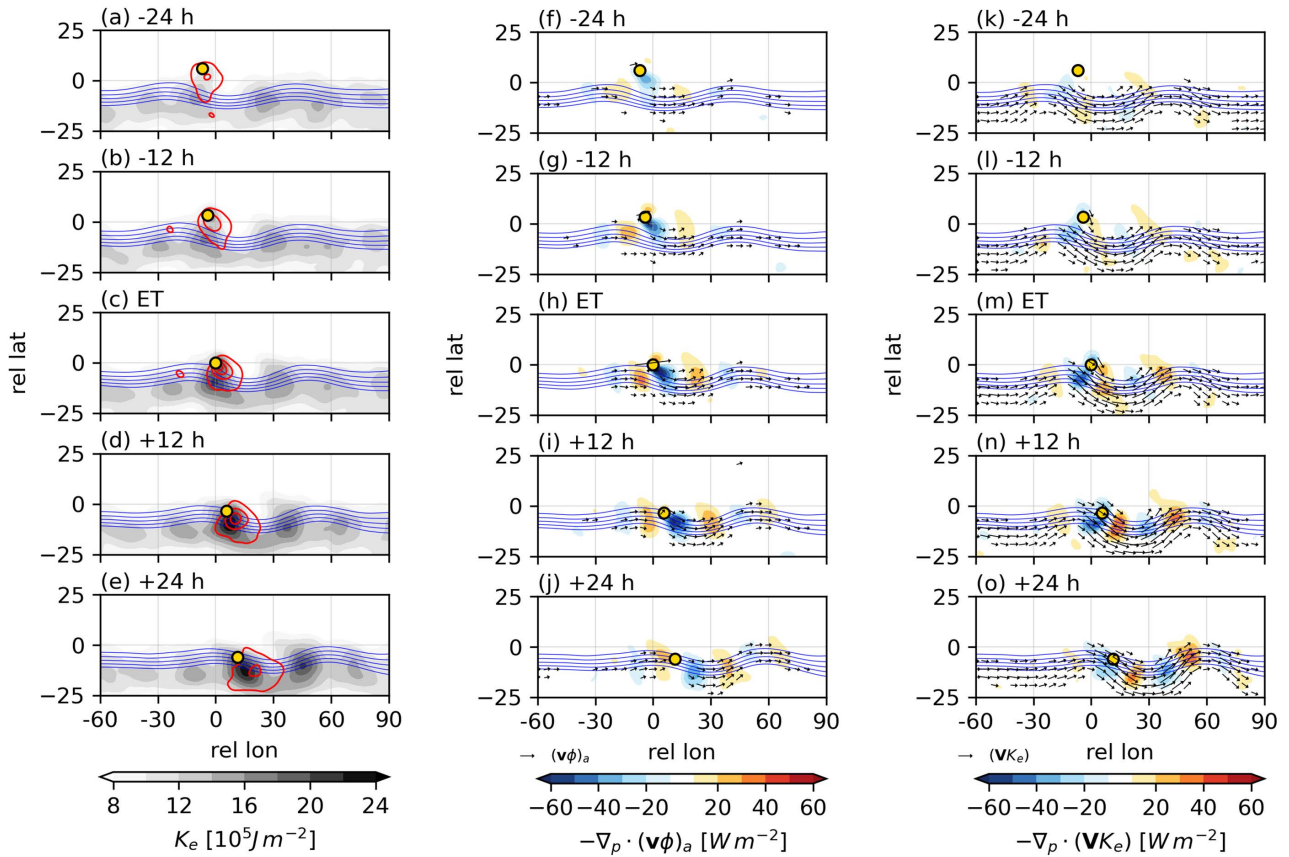


Figure 8. Lead-lag composites of K_e budget (Eq. 6) for C2 show (a-e) K_e (shaded as per the colourbar) and $bcli$ (only positive values, red contours every 10 W m^{-2} starting from 20 W m^{-2}), (f-j) ageostrophic geopotential eddy flux $ageoflux$ (vectors) and its corresponding divergence (shaded as per the colourbar), and (k-o) K_e flux (vectors) and its corresponding divergence (shaded as per the colourbar). 500-hPa geopotential height is marked by blue contours (every 5 dam starting from 550 dam) in (a-o). Yellow dots mark the cyclone position.

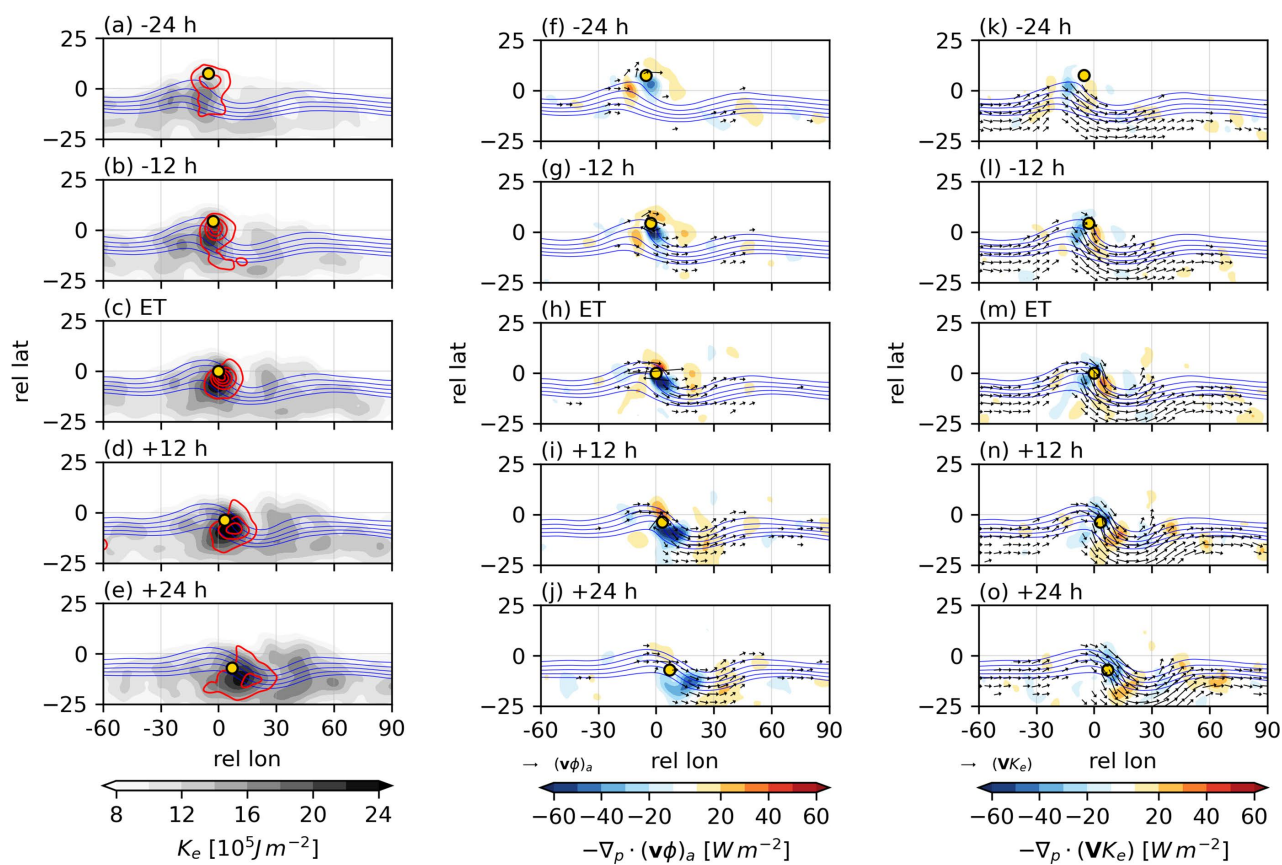


Figure 9. As in Fig. 8, but for C3.



315 3.4 Downstream influences of ET on the Australian region

We now examine the downstream influence of ET on the Australian region, with a focus of strong ET events (comprising 63 C2 events and 16 C3 events) over the southwestern Indian Ocean (SWIO; 30–90° E). For each ET event, MSLP anomalies and 330-K PV and PV anomalies are extracted from a domain centred on the TC position at ET time, which are then averaged to produce composites. Coastlines are also shown relative to this mean TC position.

320 As shown in Fig. 10a-c, a weak upstream cyclonic PV anomaly approaches the transitioning TC (indicated by negative MSLP anomalies) from its southwest side, which together propagates eastward. Immediately downstream, an upper-level ridge and its associated anticyclonic PV anomaly amplify and strengthen, accompanied by a surface anticyclone to the east of the ridge. The anticyclonic PV anomaly and positive MSLP anomaly remain strong between -24 and +24 hr, and then decay. PV contours wrap anticyclonically, and a cyclonic PV streamer forms on the eastern flank of the ridge, supporting the development
325 of a surface cyclone further downstream near the southwest of Western Australia, marked by a negative MSLP anomaly (Fig. 10a-b). This cyclone subsequently moves toward southeastern Australia (Fig. 10c-e). Meanwhile, a weak anticyclonic PV anomaly lies over the southwest of Australia at -24 hr and then propagates southeastwards ahead of the developing cyclone.

To highlight the main features of the upstream and downstream patterns of the SWIO ET events, a composite Hovmöller diagram is constructed for 350-K PV and MSLP anomalies over a fixed latitudinal band of 20–60° S (Fig. 10f). On average,
330 strong ET events over the SWIO are associated with a Rossby wave train and its corresponding alternating pattern of surface lows and highs, which persist for around 10 days and extend across approximately 180° of longitude upstream and downstream of the ET. As also indicated by Fig. 10a-c, the immediate downstream anticyclonic PV anomaly exhibits a slow propagation, particularly between -24 and +24 hr, with an estimated phase speed of 8° day^{-1} (7.8 m s^{-1}). The downstream surface cyclone and weak anticyclonic PV anomaly start to develop at around -24 hr and then last until +120 hr.

335 We then examine the frequency anomalies of heavy rainfall day (HRD) and warm day (WD) (as defined in Section 2.1) over the Australian region during and after strong SWIO ET events, and the significance of these anomalies is tested based on a Monte Carlo analysis (as described in Section 2.5). At ET, a statistically significant positive anomaly of HRD frequency concentrates in inland Western Australia (Fig. 11a), while a robust positive anomaly of WD frequency extends across the southern part of the country (Fig. 11e). After ET, HRD frequency anomalies remain weak until +72 h, when a robust positive
340 anomaly occurs across southeastern Australia (Fig. 11b–d). Meanwhile, the significant positive WD frequency anomaly shows an eastward displacement, extending over inland Western Australia to South Australia at ET and becoming confined along the eastern coastline at +72 hr (Fig. 11e–h). The eastward shift in the increased WD frequency is consistent with the progression of the weak anticyclonic PV anomaly ahead of the downstream surface cyclone (Fig. 10). As the surface cyclone approaches, the warm conditions weaken, followed by an increased frequency of HRD over the region (Fig. 11). The pattern of HRD frequency
345 anomaly at +72 h is presumably associated with a cold front extending from the cyclone. The observed synoptic patterns are typical of summertime heat events in these regions of Australia (e.g., Reeder et al., 2015).

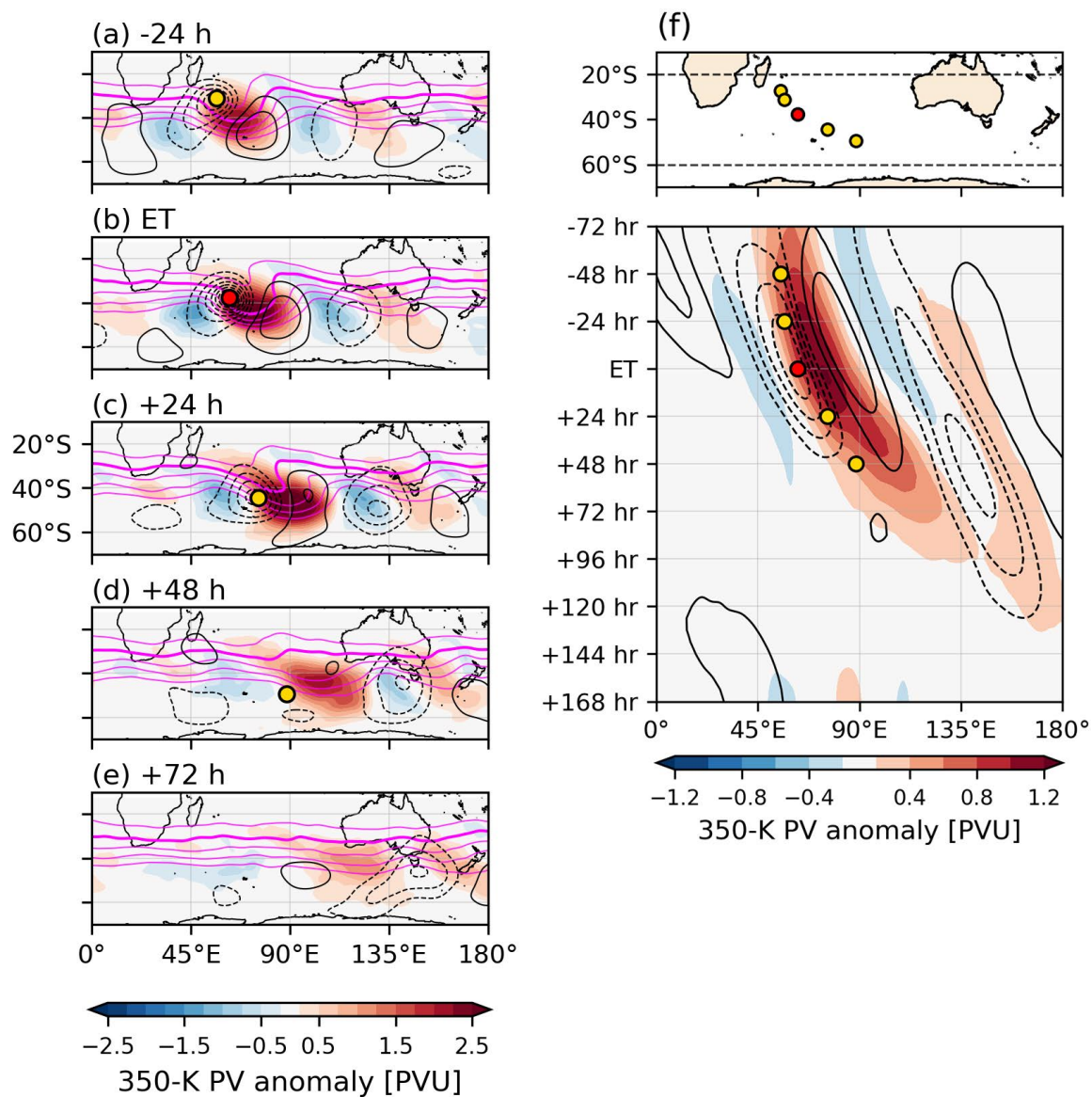


Figure 10. Lead-lag composites and composite Hovmöller diagram for C2 and C3 ET events in the southwest Indian Ocean (30–90° E). 350-K PV anomaly is shaded as per the colourbar, and MSLP anomaly is marked by black contours (every 2 hPa starting from ± 2 hPa; solid for positive and dashed for negative values) in (a-f). 350-K PV is indicated by magenta contours (every 1 PVU starting at -1 PVU), and a thick line highlights the -2 PVU contour in (a-e). Anomalies are calculated relative to the 30-day running mean. Yellow dots mark the cyclone position, and a red dot marks the cyclone position at ET in (a-e). Coastlines are also shown relative to this mean TC position. The two horizontal dashed lines in (f) mark the fixed latitude band between 20 and 60° S, over which the Hovmöller diagram is computed.

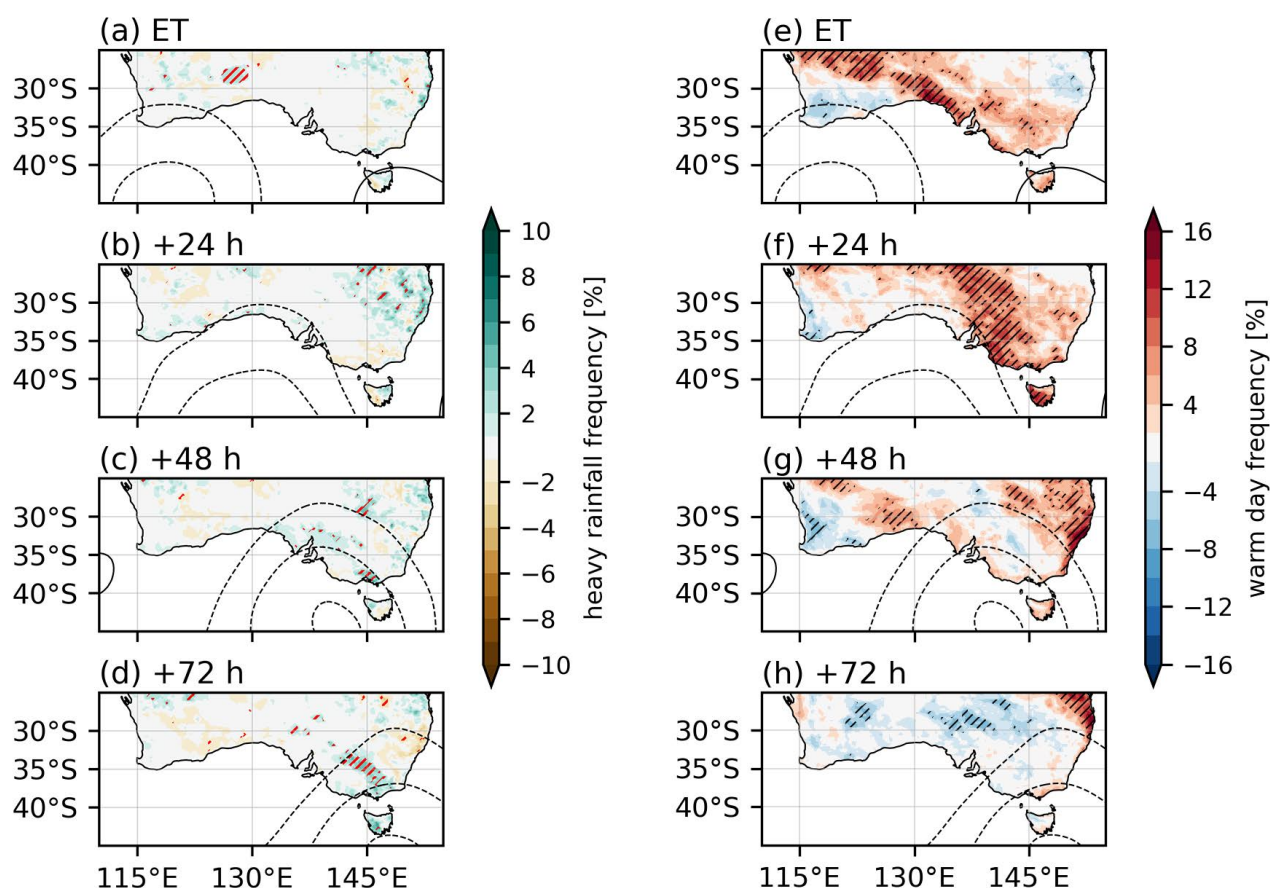


Figure 11. Frequency anomalies of (a-d) heavy rainfall days and (e-h) warm days during and after SWIO ET events over southern Australia (south of 25° S). The frequency anomalies are shaded as per the colorbar. Statistically significant anomalies are hatched in red in (a-d) and hatched in black in (e-h). MSLP anomaly is marked by black contours (every 2 hPa starting from ± 2 hPa; solid for positive and dashed for negative values).



4 Discussion

The present study diagnoses the dynamics of extratropical transition (ET) downstream development from both potential vorticity (PV) (Barnes et al., 2025) and eddy kinetic energy (K_e) (Orlanski and Sheldon, 1995) perspectives. Notably, the PV budget applied here is simpler compared to the PV framework proposed by Riemer et al. (2008), in which piecewise PV inversion is used to partition the balanced flow into components associated with different features during ET. Nevertheless, we similarly find that the divergent outflow from the TC plays a dominant role in building the downstream ridge in the early stage of ET in both clusters C2 and C3 (Figs. 6a-c and 7a-c), in line with previous studies (e.g., Riemer et al., 2008; Riemer and Jones, 2010; Archambault et al., 2013, 2015). Moreover, the nonlinear term I'_ϕ also contributes to the ridge development in C3 (Fig. 7f-j), presumably reflecting that the balanced flow associated with the tropical cyclone (TC) cyclonic PV tower advects the ridge poleward (Riemer et al., 2008; Riemer and Jones, 2010). Compared with the PV framework, the energetics diagnostic offers a more straightforward interpretation. The TC injects additional K_e by baroclinic conversion into the midlatitude flow, and the energy is then distributed by ageostrophic geopotential fluxes downstream (Figs. 8 and 9), thereby amplifying downstream Rossby wave packets (e.g., Harr and Dea, 2009; Quinting and Jones, 2016; Keller et al., 2019).

It should be noted that neither the PV nor the energetics framework directly quantifies the contribution from the diabatic processes (mainly through latent heat release). We therefore compare the vertical motion diagnosed from the Quasigeostrophic (QG) omega equation with the total vertical motion (ω), in order to evaluate the importance of dynamically forced vertical motion to that associated with diabatic processes. QG diagnostics suggest that vertical motion is substantially enhanced by diabatic processes (Figs. 4 and 5). The contribution of QG ascent to the total baroclinic conversion is likely to be smaller than that created by ascent associated with diabatic processes (Quinting and Jones, 2016), although this contribution is not explicitly computed here.

Our results identify two different pathways of downstream flow development associated with ET events over the Southern Hemisphere. In the second cluster (C2), the downstream Rossby wave train amplifies and propagates far downstream, accompanied by a succession of alternating surface lows and highs (Figs. 3a-e, 6, and 8). In contrast, C3 is characterised by a more localised downstream development, with a substantially amplified ridge immediately downstream (Figs. 3f-j, 7, and 9). This difference may be due to the distinct characteristics of the midlatitude flow in C2 and C3. In C3, the initial upper-level jet is weak, short, and poleward-oriented (Fig. 5). The weaker jet allows a large meridional deflection of the PV contours, which results in a more meridionally-oriented anticyclonic PV anomaly downstream of ET (Fig. 3f-j). With the weaker jet, the transitioning cyclone seems to be more phase-locked with the midlatitude flow during ET (Fig. 5b, e, and h), and Rossby wave packets exhibit a shorter wavelength (Fig. 7k-o). These results together explain a more pronounced and localised ridge development in C3, as shown in previous studies (Riemer et al., 2008; Riboldi et al., 2018). In addition, those jet features may also be conducive to Rossby wave breaking and hence terminate wave propagation (Wirth et al., 2018). This is further supported by a more meridionally extending downstream trough (Fig. 3h), in line with idealised simulations showing that such downstream troughs can develop when a transitioning TC interacts with a high-amplified wave pattern (Riemer and Jones, 2014). Furthermore, more meridionally-oriented K_e fluxes immediately downstream of ET (Fig. 9k-o) suggest that K_e is dispersed



away from the waveguide rather than propagating coherently downstream. Conversely, in C2, an initially broad, strong, and zonally-oriented jet (Fig. 4) favours the amplified Rossby wave packet travelling with a large group velocity and reaching far downstream (Figs. 2b, 6k-o and 8k-o). This echoes the idealised simulations of Riemer et al. (2008), in which the ET system acts as a local wave maker and promotes downstream wave development on a straight jet.

385 Although the downstream development of ET is, on average, less sensitive to the intensity and strength of the transitioning TC, these characteristics can become important for downstream impacts when the midlatitude flow already favours wave amplification (e.g., Archambault et al., 2013, 2015; Quinting and Jones, 2016; Riemer et al., 2008; Riemer and Jones, 2010). Compared with the TC in C2, the TC in C3 is initially stronger and larger (Figs. 3), which injects a higher amount of K_e downstream through baroclinic conversion at ET (Figs. 8a-e and 9a-e). Presumably, a strong and large TC can continuously
390 provide additional energy to support wave development, thereby resulting in a more amplified downstream ridge (e.g., Harr and Dea, 2009).

Our results reveal that a larger wave amplification near the ET does not guarantee a more amplified wave packet propagating farther downstream, as also highlighted by previous studies (Riemer and Jones, 2010; Pantillon et al., 2015). With the midlatitude configuration prone to wave amplification, the interaction with a strong and large TC may cause anticyclonic Rossby wave
395 breaking and interrupt the wave propagation. However, in an individual ET case (Fig. S4), Rossby wave breaks anticyclonically downstream of ET, with a PV stream formed on the east of the downstream ridge. This PV streamer subsequently stretches, wraps cyclonically, and eventually breaks up as a cut-off low. The cut-off low merges back into the midlatitude waveguide, promoting the ridge development again. This process maintains a highly-amplified ridge propagating from the southwest Indian Ocean (SWIO) to the east coast of Australia. The finding is consistent with Barnes et al. (2025), which argues that cyclonic
400 Rossby wave breaking promotes Rossby wave packet generation and downstream development.

Our results suggest that a significantly increased frequency of high-impact weather over Australia is associated with the Rossby wave promoted by the SWIO ET events (Figs. 10 and 11). Previous studies have shown that the frequency and amplitude of Rossby wave packets are enhanced downstream of the SWIO recurving TCs (Quinting and Jones, 2016). It should be noted, however, that our method does not distinguish between ET events that recurve and those that do not.

405 In general, SWIO ET events amplify the Rossby wave packet (likely through PV advection by divergent outflow I_χ) and promote associated surface lows and highs, which propagate eastward toward the Australian region (Fig. 10). The wave propagation is similar to that preceding summer heatwaves in southeastern Australia, and is identical to one of the most common propagating patterns identified in the Southern Hemisphere (O'Brien and Reeder, 2017). Nonetheless, the magnitude of the downstream anticyclonic PV anomaly in our composite is much weaker, more meridionally orientated, and more transient,
410 compared with that associated with southeastern Australia summer heatwaves (Parker et al., 2013, 2014). This suggests that SWIO ET events may contribute to warm days over Australia, but they may not necessarily drive pronounced changes in heat-wave occurrence, which requires warm conditions over consecutive days (figure not shown). Climatologically, the subtropical Indian Ocean (25–35° S) has an anticyclonic value of I_χ during November to April (Fig. S5), as TCs are generally located on the equatorward side of the subtropical jet, with their divergent outflow advecting anticyclonic PV anomaly poleward (O'Brien
415 and Reeder, 2017). Although our composites are not made relative to the point of maximum TC-midlatitude flow interaction



(measured by the I_{χ}) as did in Archambault et al. (2015), we find that most of the interaction maxima tend to occur on the southwest-south relative to the TC (Fig. S5), identical to recurving TCs over the western North Pacific.

5 Conclusions

In the present study, the dynamics of downstream development associated with individual ET clusters (as defined by (Jin et al., 2026)) are diagnosed from both PV (Barnes et al., 2025) and K_e (Orlanski and Sheldon, 1995) frameworks, with a focus on the Southern Hemisphere. Two different pathways of ET downstream development are identified. One pathway exhibits a classic picture of downstream development, with Rossby wave packets amplifying and propagating far downstream along the midlatitude waveguide. The other pathway is more localised, with a pronounced ridge immediately downstream of the transitioning TC. These distinct pathways appear to be predominantly determined by the characteristics of the midlatitude flow, although the characteristics of the TC may also play a role.

However, the current study does not quantify the relative importance of the pre-existing midlatitude flow and the TC characteristics in determining the morphology of the downstream development. Therefore, future work could address such high complexity using idealised simulations, in which the TC characteristics and background flow are systematically changed (e.g., Riemer et al., 2008; Riemer and Jones, 2010). Alternatively, PV surgery could be applied within a realistic simulation to remove the TC-related PV anomaly from the initial conditions. Therefore, the difference between the no-TC simulation and the simulation with TC can quantify the impact of the TC on downstream development (e.g., Grams et al., 2013b, a). Furthermore, the direct diabatic PV modification during ET is not explicitly considered in this study. Future studies applying passive diabatic tracers (e.g., Gray, 2006; Chagnon et al., 2013), or combining tendency diagnostics with Lagrangian trajectories (e.g., Attinger et al., 2019, 2021), could improve understanding of how diabatic processes contribute to ET downstream development. Applying these novel diagnostics to Southern Hemisphere ET events would provide new insight into the mechanisms governing downstream development.

Finally, the downstream influence of ET on high-impact weather over Australia is examined, showing a significantly increased frequency of heavy rainfall days and warm days, associated with the downstream amplification of low and high pressure circulations, respectively, over southeastern Australia after strong SWIO ET events. Future work could extend this analysis to weak ET events and to other regions in the Southern Hemisphere.



Code availability. TempestExtremes is based on code in Ullrich and Zarzycki (2017) and Ullrich et al. (2021), and it can be accessed from <https://github.com/ClimateGlobalChange/tempestextremes> (last access: 4 March 2026). The SyCLoPS software, built upon TempestExtremes, is based on the study by Han and Ullrich (2025), which can be accessed from <https://github.com/yepkids/SyCLoPS> (last access: 4 March 2026). Python code used for data analysis and visualisation in this study can be obtained upon request.

445 *Data availability.* The ERA5 reanalysis is hosted by the National Computational Infrastructure (NCI) at <https://dx.doi.org/10.25914/5f48874388857>.

Author contributions. CJ led the study and was responsible for developing the methodology, analysis, writing of the original draft and visualisation. EAR and NJH contributed to the conceptualisation, supervision and interpretation of results. MAB contributed to development of the methodology. All authors contributed to reviewing, editing and writing.

Competing interests. The authors declare that they have no conflict of interest.

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