



1 **A Dynamical Framework for Identifying Preferred Regions of Exciting Synoptic**
2 **Disturbances under Varying Background Flows**

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13 Abstract

14 Synoptic disturbances shape mid-latitude storm tracks, yet their preferred origins under
15 different background flows remain unclear. We employ a dynamical framework based on optimal
16 mode analysis to simulate synoptic disturbances and identify regions favorable to their initial
17 growth. The simulated modes reproduce key synoptic characteristics, and their root-mean-square
18 amplitude captures the spatial distribution of synoptic variability. Under Northern Hemisphere
19 climatological flow, the preferred initial region of synoptic disturbances appears over tropical–
20 subtropical Asia, while variability is stronger in December–February (DJF) than in June–August
21 (JJA). In the Southern Hemisphere, preferred initial regions shift from the South Atlantic–southern
22 Africa sector in DJF to the tropical Indian Ocean–Maritime Continent in JJA. We also examine
23 DJF background flows associated with major climate modes. During El Niño, enhanced initial
24 activity in the tropical eastern Pacific and reduced activity in the extratropical North Pacific
25 correspond to enhanced variability over southern North America and reduced variability farther
26 north, respectively. Synoptic variability dipole over the South Pacific is associated with an initial
27 meridional dipole to its west. During La Niña, Northern Hemisphere variability broadly reverses,
28 while enhanced initial activity south of Australia and New Zealand corresponds to enhanced
29 extratropical South Pacific variability. Contrasting Pacific Decadal Oscillation-related initial
30 anomalies near the Aleutians and south of Australia and New Zealand are associated with
31 downstream changes across northern mid-latitudes and the South Pacific–Southern Ocean,
32 respectively. This study identifies dynamically favorable regions for synoptic perturbation growth,
33 with potential implications for initial-condition ensemble forecasts.



34 **1 Introduction**

35 Synoptic-scale disturbances contribute substantially to mid-latitude atmospheric variability
36 and typically have timescales shorter than 10 days and wavelengths below 5000 km. Their activity
37 is closely linked to jets and waveguides (Blackmon, 1976; Lau, 1978; Chang et al., 2002; Wang
38 and Xu, 2024; Xu et al., 2024). These disturbances follow preferred pathways, or storm tracks,
39 with maxima over the North Pacific and North Atlantic in the Northern Hemisphere and a more
40 zonal band over the Atlantic and Indian Oceans in the Southern Hemisphere (Chang et al., 2002;
41 Hoskins and Hodges, 2002, 2005, 2019a,b; Shaw et al., 2022; Garfinkel et al., 2026). Global
42 warming and major modes of climate variability are expected to alter the mean atmospheric state,
43 particularly the background flow, thereby affecting the genesis and intensity of mid-latitude
44 synoptic disturbances and, consequently, weather and climate extremes (Lehmann et al., 2015;
45 Shaw et al., 2016; Tamarin-Brodsky and Kaspi, 2017; Park and Kim, 2021). Understanding where
46 synoptic disturbances initiate and how they grow under changing background flows is therefore
47 essential.

48 Synoptic disturbances in the extratropics are often characterized by eddy kinetic energy
49 (EKE). Atmospheric energy dispersion, or downstream development, is a key mechanism in their
50 evolution (Simmons and Hoskins 1979). Disturbances can grow from an upstream energy source
51 and then propagate through downstream baroclinic development (Chang, 1993, 2000; Decker and
52 Martin, 2005; Danielson et al., 2006; Rivière et al., 2015; Pinheiro et al., 2021). Local EKE budgets
53 quantify such energy dispersion in terms of baroclinic and barotropic energy conversions, energy-
54 flux convergence, and dissipation (e.g., Jiang et al., 2013; Okajima et al., 2022; Ndarana et al.,
55 2023; Zhuge et al., 2023; Battalio and Lora, 2024). Applying the EKE budget analysis, Battalio
56 and Lora (2024), for example, showed that synoptic eddy activity has increased more strongly in
57 the Southern Hemisphere over the past four decades, largely due to enhanced baroclinic conversion
58 upstream of the storm tracks, whereas changes in the Northern Hemisphere are more seasonally
59 and regionally dependent. Over the South Atlantic–southern Africa sector, Ndarana et al. (2023)
60 revealed that synoptic-scale downstream development is driven by local baroclinic conversion and
61 ageostrophic energy transfer, with cut-off lows enhancing upstream conversion and cross-ridge
62 transfer into South Africa.



63 Energetics describes the energy sources and processes associated with the “storminess” of
64 synoptic disturbances (Feser et al., 2014), whereas wave activity measures disturbance amplitude
65 and propagation, or “waviness” (Kornhuber and Messori, 2023). Using a local wave activity
66 budget analysis (Huang and Nakamura, 2016; Nakamura, 2024), Lee and Nakamura (2024)
67 estimated that cloud-related diabatic processes account for about 30% of winter wave activity over
68 the North Atlantic and North Pacific storm tracks, but estimates for the Southern Hemisphere
69 remain less certain. Other diagnostics of wave activity include the Eliassen–Palm flux (Edmon et
70 al., 1980), Plumb wave activity flux (Plumb, 1985), Takaya–Nakamura wave activity flux (Takaya
71 and Nakamura, 2001), and Rossby wave source (Senapati et al., 2024). Lagrangian methods,
72 sometimes combined with Eulerian diagnostics, track individual synoptic disturbances and their
73 origins. Berry et al. (2012), for instance, tracked coherent maxima of isentropic potential vorticity
74 associated with synoptic disturbances and found that most disturbances originate in the Australian
75 monsoon region. A Rossby wave ray flux approach combines ray tracing with Eulerian flux
76 characteristics to describe propagation pathways and intensity of Rossby wave activity (Yang and
77 Li, 2025; Li and Li, 2026). Moist wave activity measures the area-integrated meridional
78 displacement of column water vapor contours from their equivalent zonal contours, characterizing
79 large-scale moist intrusion and transport (Xue et al., 2026).

80 These approaches mainly diagnose the energy, wave activity, and propagation of identified
81 synoptic disturbances. They do not directly isolate regions where weak initial perturbations
82 preferentially grow under a specified background flow or show how changes in that flow alter their
83 growth. Dynamical instability analysis addresses this question by examining perturbation growth
84 about a prescribed basic state. Growth may be modal, transient, or optimal (Mak 2011). In modal
85 growth, the most unstable mode eventually dominates (e.g., Hall and Sardeshmukh, 1998; Yang
86 and Li, 2023), whereas transient growth describes rapid initial amplification (Mamatsashvili et al.,
87 2010). Optimal growth selects the initial perturbation that achieves the largest amplification within
88 a specified time interval (Chang and Mak, 1995). Because synoptic disturbances typically have
89 lifetimes of less than 10 days, their development is better characterized within a specified time
90 interval in the optimal growth framework. Within this framework, candidate initial growth regions
91 can be defined as locations where leading optimal perturbations are initially concentrated (Zhao et
92 al., 2020, 2026).



93 Optimal growth has been widely studied in baroclinic wave excitation, explosive
 94 cyclogenesis, intraseasonal variability, and even El Niño–Southern Oscillation (ENSO) dynamics
 95 (e.g., Farrell, 1989; Winkler et al., 2001; Mitas and Robinson, 2005; Newman et al., 2011; Vimont
 96 et al., 2014; Zhao et al., 2018, 2020; Mak et al., 2021). Storm track dynamics has also been
 97 investigated using optimal mode analysis. For example, earlier studies identified amplifying
 98 structures over East Asian, North Atlantic–European, and North American storm track regions
 99 (e.g., Lee and Mak, 1995). More recent work has used the root-mean-square (RMS) amplitude and
 100 kernel density of multiple optimal modes to assess how background flow changes affect synoptic
 101 variability. Zhao et al. (2020) related the recent decline in northern mid-latitude summer variability
 102 to reduced variance among fast-growing optimal modes. Zhao et al. (2026) found that marine cloud
 103 brightening over tropical and subtropical oceans can alter winter–early spring background flow,
 104 suppress synoptic disturbance growth over mid-eastern North America, and contribute to
 105 unintended regional warming.

106 What remains unclear is how climatological flows and flows associated with global climate
 107 modes, including the ENSO and Pacific Decadal Oscillation (PDO), shape intrinsic synoptic
 108 growth in both hemispheres. Studies of global climate variability have examined, for instance,
 109 eddy–mean flow feedbacks associated with the Southern Annular Mode (e.g., Nie et al., 2013;
 110 Thompson and Li, 2015) and downstream eddy development linking the ENSO to North Atlantic
 111 Oscillation (e.g., Li and Lau, 2012). They have focused less on the preferred regions for exciting
 112 synoptic disturbances under these flows. Here, we use optimal mode analysis to assess whether
 113 the RMS of the optimal modes under different background flows resembles the spatial patterns of
 114 observed synoptic variability in both hemispheres. We then identify regions favorable to initial
 115 disturbance growth and examine how those regions change under different background flows.
 116 Section 2 describes the data and methods. Section 3 first examines disturbance pathways and
 117 preferred initial regions under climatological flows, and then assesses their responses to global
 118 climate modes. Sections 4 and 5 present the conclusions and discussion, respectively.

119 **2 Data and methodology**

120 **2.1 Data and event selection**

121 We obtain daily mean zonal and meridional winds and geopotential height from the European
 122 Centre for Medium-Range Weather Forecasts (ECMWF) fifth-generation reanalysis (ERA5;



Hersbach et al., 2020) for 1979–2024. We use a $2.5^\circ \times 2.5^\circ$ horizontal grid to focus on large-scale atmospheric dynamics, and the results are insensitive to horizontal resolution. We use the Oceanic Niño Index (ONI) provided by the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC), defined as the three-month running mean sea surface temperature (SST) anomaly in the Niño-3.4 region. Following the CPC classification, an El Niño event is identified when the ONI reaches or exceeds $+0.5^\circ\text{C}$ for at least five consecutive overlapping three-month periods. Based on the CPC ONI record, strong El Niño conditions ($\text{ONI} > 1.5^\circ\text{C}$) occurred in DJF 1982/83, 1987/88, 1991/92, 1997/98, 2015/16, and 2023/24, and these seasons are used for the El Niño composite. Strong La Niña conditions ($\text{ONI} < -1.5^\circ\text{C}$) occurred in DJF 1988/89, 1998/99, 1999/2000, 2007/08, and 2010/11, and these seasons are used for the La Niña composite. We identify PDO phases using the Extended Reconstructed Sea Surface Temperature (ERSST) PDO Index provided by the NOAA National Centers for Environmental Information. Rather than using the entire duration of each PDO phase, we select five-year periods with relatively large PDO magnitudes to enhance the contrast in background flows between each phase and the climatology. Given the general shift from positive to negative PDO values since 1979, DJF 1982/83–1986/87 and DJF 2019/20–2023/24 are selected to represent the positive and negative PDO phases, respectively, and a composite is calculated for each period.

2.2 Dynamical framework based on optimal mode analysis

Following Mak (2011) and Zhao et al. (2020, 2026), we use a baroclinic model based on the three-level quasi-geostrophic (QG) potential vorticity equations (Step 1 in Fig. 1). The equations for the upper and lower levels, with the middle level associated with the parameterization of static stability parameter, are:

$$\begin{aligned} \frac{\partial}{\partial t} \left[\zeta_1 + \frac{f(Z_3 - Z_1)}{(\delta p)^2 S_2} \right] + f^{-1} J(Z_1, \zeta_1 + f) + \frac{1}{(\delta p)^2 S_2} J(Z_1, Z_3) &= -\kappa \zeta_1 \\ \frac{\partial}{\partial t} \left[\zeta_3 + \frac{f(Z_1 - Z_3)}{(\delta p)^2 S_2} \right] + f^{-1} J(Z_3, \zeta_3 + f) + \frac{1}{(\delta p)^2 S_2} J(Z_3, Z_1) &= -\kappa \zeta_3. \end{aligned} \quad (1)$$

In Eq. (1), subscripts 1, 2, and 3 are 200 hPa, 400 hPa, and 600 hPa levels, respectively. ζ and Z are the geostrophic vorticity and geopotential, respectively, with a relation of $\zeta = f^{-1} \mathcal{L}\{Z\}$ ($\mathcal{L}\{\}$ is the Laplacian operator); J is the Jacobian operator. The Coriolis frequency, layer thickness, dissipation coefficient, and static stability parameter are denoted by f , δp , κ , and S , respectively.



149 Static stability parameter at 400 hPa (S_2) is prescribed as $2.1 \times 10^{-6} \text{ m}^2 \text{ Pa}^{-2} \text{ s}^{-2}$ for boreal
150 winter and $3.1 \times 10^{-6} \text{ m}^2 \text{ Pa}^{-2} \text{ s}^{-2}$ for summer (Zhao et al., 2020, 2026). To represent
151 topographic and land–sea contrast, we parameterize the dissipation coefficient (κ), with the
152 damping timescale (κ^{-1}) of 5 days at 200 hPa over both ocean and land, and 4 days over the ocean
153 and less than 3 days over land at 600 hPa.

154 We linearize Eq. (1) about long-term seasonal mean geopotential $\bar{Z}$ (Step 2 in Fig. 1),
155 obtained by multiplying seasonal-mean geopotential height by gravitational acceleration. The
156 linearized equations are:

$$\begin{aligned}
 157 \quad & \frac{\partial}{\partial t} \left[\zeta'_1 + \frac{f}{(\delta p)^2 S_2} (Z'_3 - Z'_1) \right] + \frac{1}{f} \left(Z'_{1x} \bar{\zeta}_{1y} + \bar{Z}_{1x} \zeta'_{1y} + Z'_{1x} f_y - Z'_{1y} \bar{\zeta}_{1x} - \bar{Z}_{1y} \zeta'_{1x} \right) \\
 158 \quad & + \frac{1}{S_2 (\delta p)^2} (\bar{Z}_{1x} Z'_{3y} + Z'_{1x} \bar{Z}_{3y} - \bar{Z}_{1y} Z'_{3x} - Z'_{1y} \bar{Z}_{3x}) = -\kappa \zeta'_1 \\
 159 \quad & \frac{\partial}{\partial t} \left[\zeta'_3 + \frac{f}{(\delta p)^2 S_2} (Z'_1 - Z'_3) \right] + \frac{1}{f} \left(Z'_{3x} \bar{\zeta}_{3y} + \bar{Z}_{3x} \zeta'_{3y} + Z'_{3x} f_y - Z'_{3y} \bar{\zeta}_{3x} - \bar{Z}_{3y} \zeta'_{3x} \right) \\
 160 \quad & + \frac{1}{S_2 (\delta p)^2} (\bar{Z}_{3x} Z'_{1y} + Z'_{3x} \bar{Z}_{1y} - \bar{Z}_{3y} Z'_{1x} - Z'_{3y} \bar{Z}_{1x}) = -\kappa \zeta'_3. \quad (2)
 \end{aligned}$$

161 In Eq. (2), the overbar and prime denote the seasonal mean and perturbation, respectively. The
162 subscripts x and y represent derivatives with respect to longitude and latitude, respectively. ζ can
163 be further expressed by $f^{-1} \mathcal{L}\{Z\}$. We then assume solutions for the upper- and lower-level
164 geopotential perturbation of the form:

$$\begin{aligned}
 Z'_1 &= \Phi_1 e^{i\sigma t} \\
 Z'_3 &= \Phi_3 e^{i\sigma t}, \quad (3)
 \end{aligned}$$

165 where Φ , σ , and i denote the perturbation amplitude, oscillation frequency and imaginary unit,
166 respectively. We substitute Eq. (3) into Eq. (2). Because the domain is primarily extratropical
167 rather than global, we use second-order centered finite differences, with cyclic boundary
168 conditions in the longitudinal direction applied. This yields the matrix equation:

$$\sigma \mathbf{L} \begin{pmatrix} \Phi_1 \\ \Phi_3 \end{pmatrix} = \mathbf{R} \begin{pmatrix} \Phi_1 \\ \Phi_3 \end{pmatrix}, \quad (4)$$



169 where $\mathbf{L}$ and $\mathbf{R}$ are matrices ($2N \times 2N$; N is the total grid points in the selected two-dimensional
 170 gridded domain). We solve the eigenvalue problem in Eq. (4), yielding $2N$ compound normal
 171 modes, each consisting of coupled upper- and lower-level components.

172 We use these normal modes as base functions to construct optimal modes over a prescribed
 173 optimization time τ . Here, τ is set to 2 days, approximately one-quarter of the typical lifetime of
 174 synoptic disturbances. The geopotential perturbation ($\mathbf{Z}$) at $t = \tau$ over all grid points in the domain
 175 at the upper and lower levels can be expressed as:

$$\mathbf{Z} = \sum_{n=1}^{2N} \begin{pmatrix} \Phi_1 \\ \Phi_3 \end{pmatrix}_n e^{i\sigma_n \tau} a_n, \quad (5)$$

176 where a_n is the projection coefficient for each normal mode. Thus, Eq. (5) represents the
 177 geopotential perturbation as a linear combination of the normal modes, with different sets of
 178 projection coefficients a_n defining different combinations of the normal modes. We measure
 179 disturbance growth with an energy norm (e.g., Mitas and Robinson 2005; Mak 2011):

$$A(\tau) = \mathbf{Z}^H \mathbf{D} \mathbf{Z}, \quad (6)$$

180 where $\mathbf{Z}^H$ is the Hermitian of $\mathbf{Z}$ and $\mathbf{D}$ is a unit matrix. We define the amplification factor γ as the
 181 ratio of disturbance energy at $t = \tau$ day to that at $t = 0$ day:

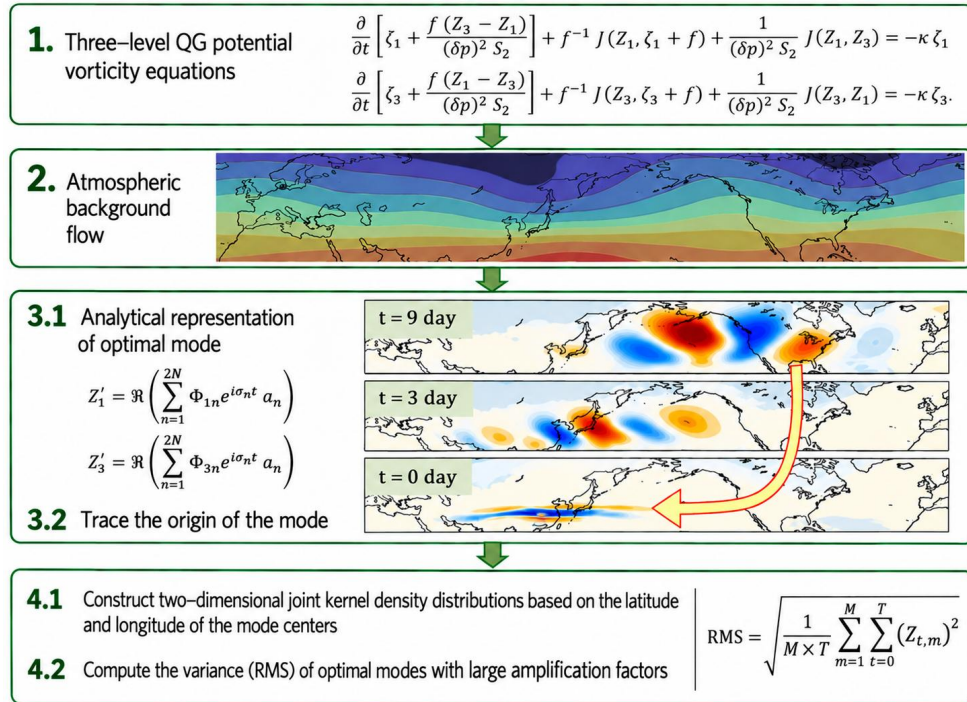
$$\gamma \equiv \frac{A(\tau)}{A(0)}. \quad (7)$$

182 Solving the eigenvalue problem for the disturbance energy (A), with γ as eigenvalue and $\mathbf{a}$ as the
 183 eigenvector (the vector of the elements $\{a_n\}$), yields $2N$ optimal modes associated with pairs of
 184 γ and $\mathbf{a}$. The analytical representation of the optimal modes at the upper and lower levels (Step
 185 3.1 in Fig. 1) can then be computed from:

$$\begin{aligned} \mathbf{Z}_1 &= \Re \left(\sum_{n=1}^{2N} \Phi_{1n} e^{i\sigma_n t} a_n \right) \\ \mathbf{Z}_3 &= \Re \left(\sum_{n=1}^{2N} \Phi_{3n} e^{i\sigma_n t} a_n \right). \end{aligned} \quad (8)$$



186 Each component in Eq. (8) can be obtained from the solution of the eigenvalue–eigenvector
 187 problem, with the real part $\Re$ used for the final result.



188
 189 **Figure 1.** Framework and objectives. Step 1: Derive the three-level QG potential vorticity
 190 equations. Step 2: Prescribe the atmospheric background flow. Step 3: Solve the linearized QG
 191 potential vorticity equations to obtain the analytical representation of the optimal modes and trace
 192 their origins by examining their temporal evolution. The schematic illustrates the backward tracing
 193 of the initial disturbance associated with an optimal mode. Step 4: Construct two-dimensional joint
 194 kernel density distributions of optimal mode centers and examine synoptic variability using the
 195 RMS of the optimal modes with large amplification factors.

196

197 We locate each mode daily over its 10-day evolution ($t = 0 - 9$ days). The initial location of
 198 each mode is determined by setting $t = 0$ (see the schematic in Step 3.2 of Fig. 1), allowing us to
 199 locate the mode's original location. For each day, the mode center is the grid point of maximum
 200 amplitude. We construct two-dimensional joint kernel densities of these centers (Step 4.1 in Fig.
 201 1). We further examine the variability represented by the optimal modes by calculating the RMS



202 amplitude of the first M modes with largest amplification factors over a specified period of their
 203 evolution (from $t = 0$ to $t = T$ day) (Step 4.2 in Fig. 1). Before calculating the RMS, each mode
 204 is normalized by its maximum amplitude over the entire 10-day evolution; that is, its amplitude at
 205 each grid point and time step is divided by this maximum value. The RMS is expressed as:

$$206 \quad \text{RMS} = \sqrt{\frac{1}{M \times T} \sum_{m=1}^M \sum_{t=0}^T (\mathbf{z}_{t,m})^2} \quad (9)$$

207 When the selected period T is confined to the early stage of mode evolution (e.g., day 0–1), the
 208 RMS characterizes the spatial distribution of initial-stage disturbance amplitude, with large values
 209 indicating regions where perturbations preferentially initiate. Because each mode is normalized by
 210 its own maximum over the 10-day evolution, the resulting RMS is dimensionless. Its comparison
 211 with observed EKE concerns spatial patterns, while their magnitudes are not directly comparable.

212 We test the sensitivity of optimal mode analysis. Adding or removing a few years has little
 213 effect on the long-term climatological seasonal-mean flow, while the flows associated with climate
 214 modes are more sensitive to year selection. The years selected in Section 2.1 ensure that the
 215 corresponding background flows are distinct from the climatology. Joint kernel densities, RMS
 216 patterns, and amplification factors change little when we vary the number of leading modes,
 217 including tests with 10, 50, and 100 modes. Results are also similar at 300, 500, and 700 hPa; with
 218 dissipation timescales shifted by several days; and with optimization periods of 1–5 days.

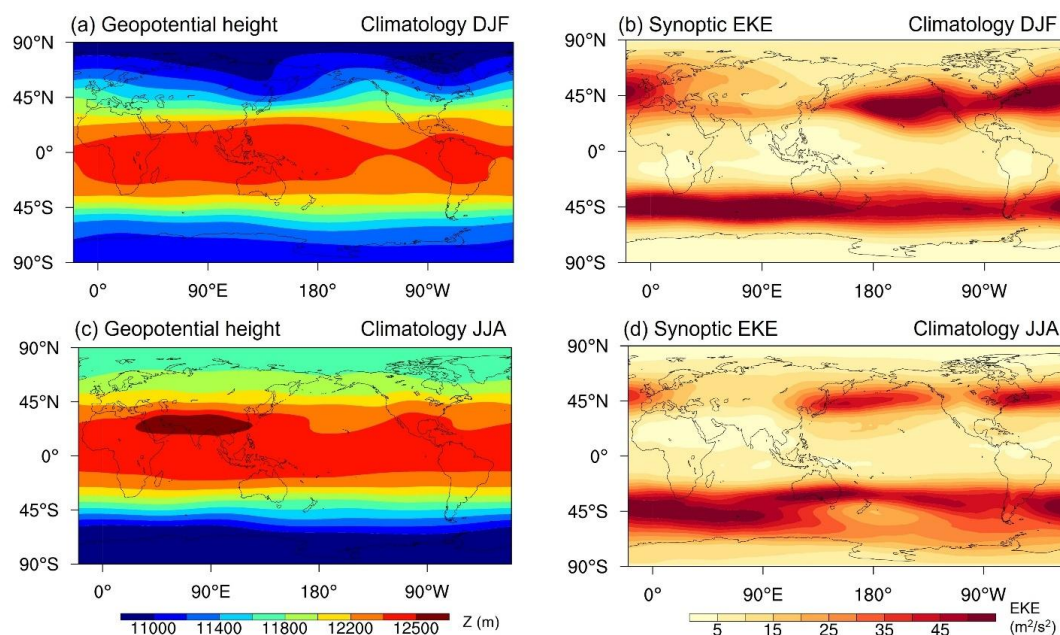
219 **3. Results**

220 **3.1 Preferred synoptic disturbance pathways and origins under climatological flows**

221 In DJF, climatological 200-hPa geopotential height generally decreases toward the poles in
 222 both hemispheres, with prominent troughs over high-latitude Asia and North America (Fig. 2a).
 223 Synoptic EKE is defined as: $(u'^2 + v'^2)/2$, where u' and v' are 2.5–6 day Butterworth band-pass
 224 filtered (Russell, 2006) 200 hPa zonal and meridional winds, respectively. It is large over the North
 225 Pacific and North Atlantic storm tracks and forms a zonal band near 45°S (Fig. 2b), consistent
 226 with previous work (e.g., Chang et al., 2002; Hoskins and Hodges, 2002, 2005, 2019a,b). In JJA,
 227 geopotential height increases toward northern high latitudes but decreases toward southern high
 228 latitudes; the South Asian High dominates subtropical Asia (Fig. 2c). Compared to DJF, Northern



229 Hemisphere synoptic EKE weakens and the strong EKE region south of Australia shifts
 230 equatorward (Fig. 2d). We next examine disturbance growth under these seasonal-mean
 231 background flows.



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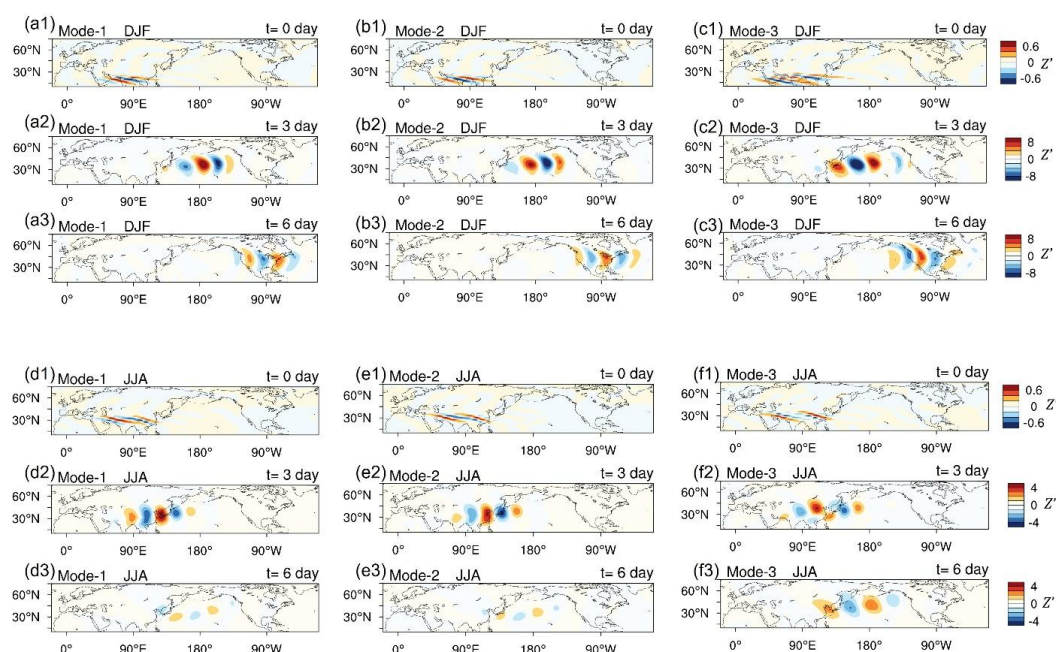
233 **Figure 2.** (a) Climatological geopotential height (units: m) at 200 hPa during DJF averaged over
 234 1979–2024 in reanalysis. (b) As in (a), but for synoptic EKE (unit: $\text{m}^2 \text{s}^{-2}$). (c–d) As in a–b, but
 235 for JJA.

236

237 We apply optimal mode analysis to the climatological flows to identify synoptic disturbances
 238 that can grow within them. The analysis does not reproduce the timing or precise locations of
 239 disturbances in reanalysis, but it examines how each background flow favors perturbation growth
 240 and propagation (Zhao et al., 2026). Extending previous studies (Zhao et al., 2020, 2026), we
 241 examine climatological DJF and JJA flows in both hemispheres. As described in Section 2.2, we
 242 examine disturbances that can grow rapidly within a 2-day time interval under climatological flow.
 243 Figure 3a–c shows the evolution of the first three optimal modes at 0, 3, and 6 days, respectively,
 244 under the climatological Northern Hemisphere background flow during DJF. These modes have
 245 the three largest amplification factors (γ) and thus represent the fastest-growing upper-



246 tropospheric synoptic-scale eddies. They exhibit similar propagation characteristics, originating
 247 over tropical and subtropical Central–East Asia south of 30°N and propagating eastward across
 248 the North Pacific toward North America (Fig. 3a1–c1), with slight differences in phase. Their
 249 wavelengths are approximately 40° in longitude (~4000 km), consistent with the typical scale of
 250 synoptic disturbances. During JJA, the optimal modes initiate farther north, over subtropical
 251 Central–East Asia near 30°N (Fig. 3d1–f1), similar to the preferred initiation region of
 252 intraseasonal disturbances under the JJA background flow (Zhao et al., 2018). Compared with DJF,
 253 these disturbances develop more slowly and dissipate over the North Pacific by day 6 (Fig. 3d–f).



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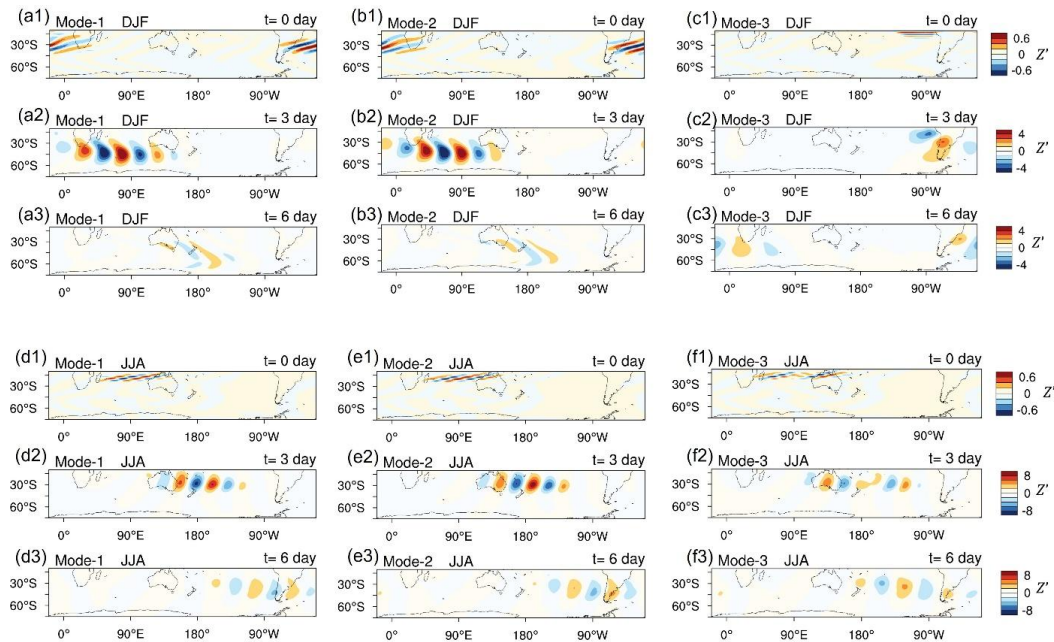
255 **Figure 3. (a1–a3)** Time evolution ($t = 0, 3$, and 6 days) of the 200 hPa geopotential perturbations
 256 (Z' ; non-dimensional) associated with the fastest-growing synoptic-scale eddy (optimal mode 1)
 257 during DJF in the climatological Northern Hemisphere background flow. Note that panel **a1** uses
 258 a different color bar from panels **a2** and **a3**. **(b1–b3 and c1–c3)** As in **(a1–a3)**, but for the second-
 259 and third-fastest-growing eddies (optimal modes 2 and 3), respectively. **(d–f)** As in **(a–c)**, but for
 260 JJA.

261

262 In the Southern Hemisphere during DJF, the first two modes originate in the South America–
 263 South Atlantic–southern Africa sector and propagate southeastward across the Indian Ocean



264 toward the South Pacific, with slight phase differences (Fig. 4a–b). The third starts over the tropical
 265 South Pacific and propagates southeastward across South America toward the South Atlantic and
 266 Indian Ocean (Fig. 4c), consistent with a documented tropical-to-extratropical Rossby-wave
 267 pathway (Li et al., 2019). In JJA, the first three modes originate over the tropical Indian Ocean and
 268 Maritime Continent and move eastward across the Pacific, reaching South America around day 6
 269 (Fig. 4d–f). Thus, the leading modes reveal seasonally varying regions and pathways favorable to
 270 disturbance growth in both hemispheres.



271

272 **Figure 4. (a–f)** As in Fig. 3, but for the Southern Hemisphere.

273

274 To evaluate synoptic variability, we examine the 50 modes with large amplification factors
 275 (Zhao et al., 2020, 2026). We calculate the joint kernel density of their centers in latitude–longitude
 276 space for each background flow (see Section 2.2). Under Northern Hemisphere DJF climatological
 277 flow, mode centers dominate over the northern mid-latitudes (Fig. 5a). Furthermore, earlier studies
 278 have suggested that the variance of transients from multiple short-term integrations of a linear
 279 storm-track model can be used to estimate synoptic variability (e.g., Branstator, 1995; Gritsun et



al., 2008). Adopting this approach, we treat each optimal mode as an equally weighted model solution and calculate the RMS of the leading 50 modes as a spatial proxy for synoptic variability (see Section 2.2).

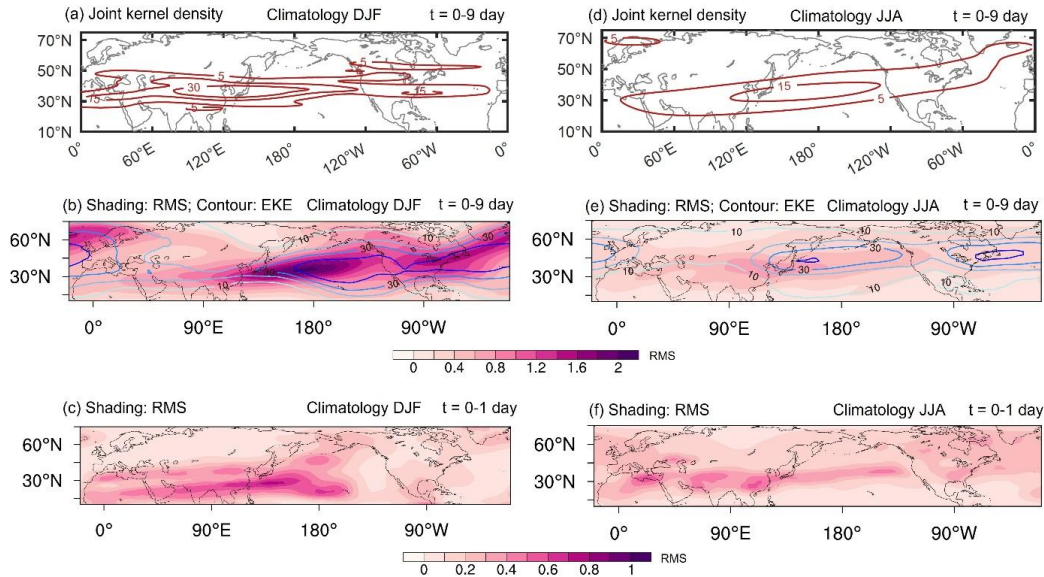
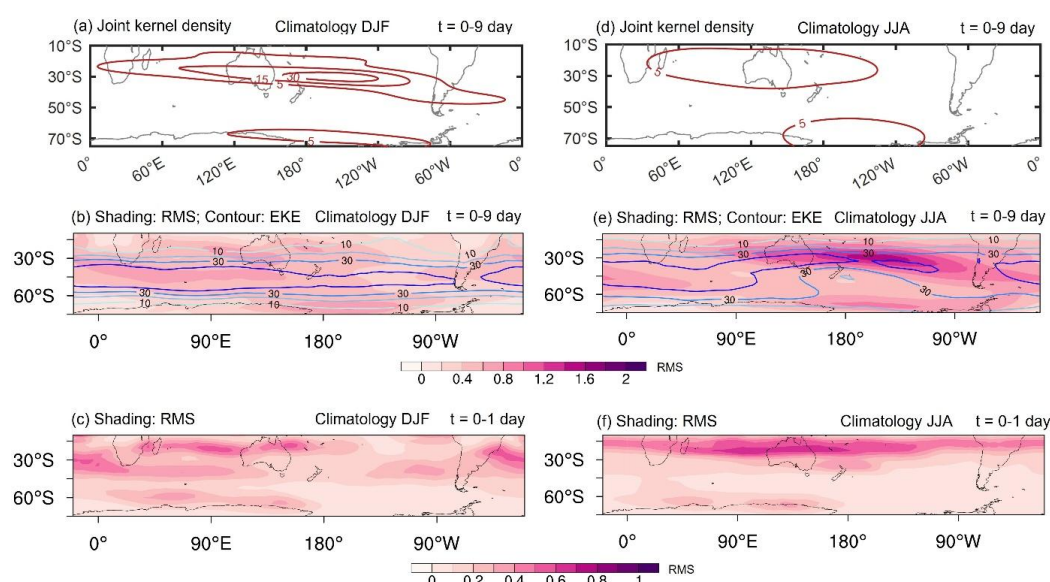


Figure 5. (a) Joint kernel density distributions ($\times 10^{-5}$) of the occurrence locations of the first 50 optimal modes, defined as the locations of their maximum amplitudes, during their 10-day evolution ($t = 0-9$ days) in the Northern Hemisphere climatological background flow during DJF. (b) The RMS of geopotential perturbations (shading; non-dimensional) of the first 50 optimal modes during their 10-day evolution ($t = 0-9$ days). Contours show climatological synoptic EKE (unit: $\text{m}^2 \text{s}^{-2}$) at 200 hPa during DJF in reanalysis. The contour ranges from 10 to 45 (unit: $\text{m}^2 \text{s}^{-2}$). (c) Shading is same as in (b), but for the initial stage ($t = 0-1$ day). Note that panel c uses a different color bar from panel b. (d-f) As in (a-c), but for JJA.

Over the 10-day period in DJF, high RMS amplitudes occur over the North Pacific and the eastern North America–North Atlantic sector (Fig. 5b), spatially consistent with observed climatological synoptic EKE (Fig. 2b; contours in Fig. 5b). This agreement indicates broad geographic similarity between the optimal modes' RMS and observed EKE patterns. During the initial stage ($t = 0-1$ day), RMS is largest over tropical–subtropical Asia and the western North Pacific (Fig. 5c), encompassing the initial locations of the three leading modes (Fig. 3a1–c1). In JJA, mode-center density remains high over northern mid-latitude storm-track regions (Fig. 5d),



while the 10-day RMS weakens (Fig. 5e) compared to DJF. Initial RMS maxima appear near 30°N from North Africa across subtropical Asia to the North Pacific (Fig. 5f), consistent with the initial locations of the three leading modes (Fig. 3d1–f1). Together, Figs. 3 and 5 show how initially concentrated synoptic perturbations develop and propagate downstream toward regions of maximum amplitude.



305

306 **Figure 6. (a–f)** As in Fig. 5, but for the Southern Hemisphere.

307

308 In the Southern Hemisphere during DJF, the centers of the first 50 optimal modes appear over
 309 Australia and the subtropical to mid-latitude South Pacific (Fig. 6a). Their 10-day RMS has no
 310 pronounced maximum, although values are relatively high over the mid-latitude Indian Ocean
 311 (shading in Fig. 6b), broadly consistent with the climatological EKE belt. (contours in Fig. 6b).
 312 Initial RMS is high from the South Atlantic and southern Africa across the Indian Ocean to the
 313 South Pacific (Fig. 6c), consistent with the first two DJF modes (Fig. 4a1–b1). In JJA, mode
 314 centers dominate over the tropical and subtropical Indian Ocean, Australia, and subtropical South
 315 Pacific, with another center at higher South Pacific latitudes (Fig. 6d). The 10-day RMS peaks
 316 over the subtropical South Pacific and closely matches climatological EKE (Fig. 6e). Initial RMS



317 maxima extend across the tropical Indian Ocean, Australia, and South Pacific (Fig. 6f), consistent
 318 with the three leading JJA modes (Fig. 4d1–f1).

319 To summarize, the normalized optimal modes' RMS broadly resembles the observed
 320 climatological storm-track EKE patterns in both hemispheres and seasons. Initial-stage RMS
 321 identifies regions favorable for disturbance growth, generally equatorward of the mid-latitude
 322 storm tracks, indicating that synoptic variability primarily originates in the tropical and subtropical
 323 regions before developing and propagating into the mid-latitudes.

324 **3.2 Synoptic variability and its origin associated with global climate modes**

325 Because ENSO has a stronger influence on atmospheric circulations in DJF than in JJA, we
 326 focus on DJF when comparing climate-mode background flows. We first examine ENSO
 327 composites (see Section 2.1). During El Niño, geopotential height anomalies are positive over the
 328 tropical Pacific and negative over the extratropical Pacific, especially in the Northern Hemisphere
 329 (Fig. 7a). This pattern is consistent with a Gill-type response to tropical Pacific warming followed
 330 by Rossby-wave responses (Gill, 1980; Zebiak, 1986). La Niña produces a generally opposite
 331 pattern over the tropical and extratropical Pacific (Fig. 7b).

332 These background-flow differences coincide with distinct synoptic EKE anomalies. During
 333 El Niño, EKE decreases over northwestern North America and increases farther south (Fig. 7c),
 334 indicating reduced and enhanced synoptic eddy activity in these regions, respectively. This pattern
 335 is consistent with reports of fewer cyclone landfalls in the northwest and more in the south (Jiang
 336 and Deng, 2011; Jiang et al., 2014; Zhao et al., 2017). La Niña broadly reverses this contrast (Fig.
 337 7d). In addition, El Niño EKE anomalies are also positive over Eurasia and northern Africa (Fig.
 338 7c), whereas La Niña anomalies are negative farther equatorward (Fig. 7d). In the Southern
 339 Hemisphere, El Niño produces positive anomalies over the southern Indian Ocean and
 340 southeastern Pacific, while La Niña generally reverses these and produces positive bands in parts
 341 of the Southern Ocean (Fig. 7c–d). Thus, synoptic EKE anomalies reverse between ENSO phases,
 342 with some asymmetry.

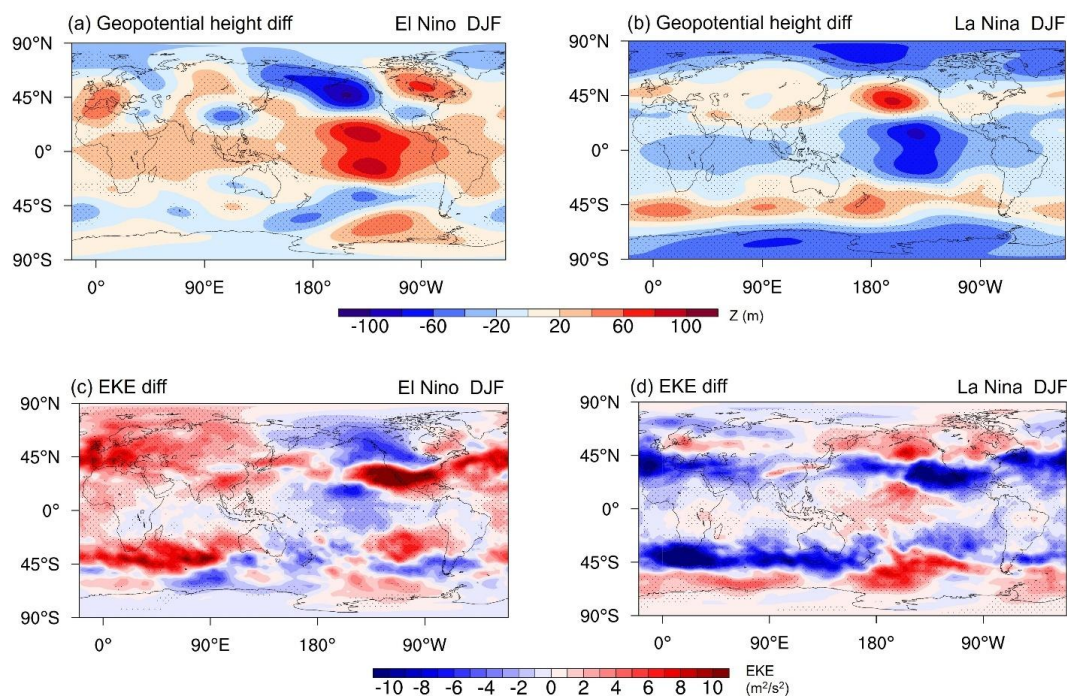


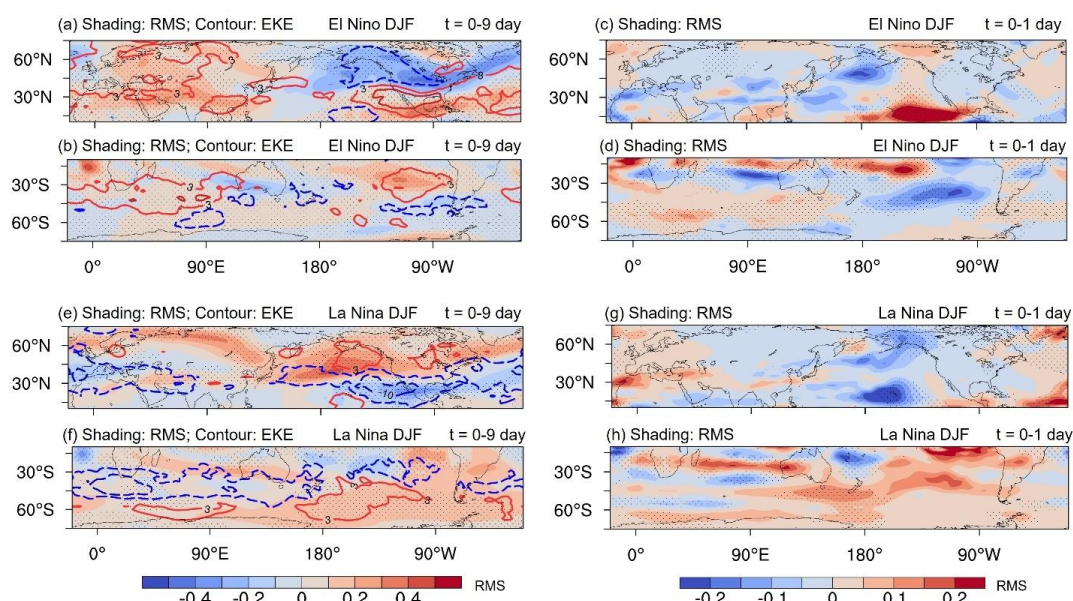
Figure 7. (a) Composite geopotential height (units: m) anomalies (relative to climatology) at 200 hPa during DJF for El Niño years in reanalysis. Stippling indicates regions where geopotential height anomalies are significant at the 5% level based on the two-tailed Student's *t*-test. (b) As in (a), but for La Niña years. (c–d) As in (a–b), but for synoptic EKE (unit: $\text{m}^2 \text{s}^{-2}$).

We next compare these EKE anomalies with optimal-mode RMS anomalies. During El Niño, the 10-day RMS shows a north–south contrast over North America and positive anomalies over Eurasia and northern Africa (shading in Fig. 8a), matching observed EKE (Fig. 7c; contours in Fig. 8a). In the Southern Hemisphere, optimal mode analysis captures the eastern extratropical South Pacific contrast but underestimates positive variability over the southern Indian Ocean (Fig. 8b). At the initial stage, RMS decreases relative to climatology over the extratropical North Pacific and increases over the tropical eastern Pacific (Fig. 8c), while a meridional dipole appears over the South Pacific (Fig. 8d). Note that these initial-stage anomalies indicate departures from the climatological RMS, rather than the absolute distribution of synoptic variability under El Niño. Accordingly, the positive anomaly over the tropical eastern Pacific does not identify this region as a principal origin of the leading optimal modes (figure not shown), but instead indicates enhanced



360 local synoptic variability relative to climatology, possibly associated with stronger or more
 361 frequent disturbances present in or passing through the region at this stage. These enhanced initial
 362 disturbances (Fig. 8c–d) may subsequently influence southern North America and the subtropical
 363 South Pacific downstream, with possible further connections to Eurasia, northern Africa, and the
 364 southern Indian Ocean (Fig. 8a–b). Reduced initial RMS over the North Pacific (Fig. 8c) may
 365 likewise contribute to weaker variability over northern North America and the North Atlantic
 366 storm track (Fig. 8a).

367 During La Niña, 10-day RMS anomalies over the Pacific–America sector broadly reverse
 368 compared to El Niño (shading in Fig. 8e–f), consistent with observed EKE anomalies. At the initial
 369 stage, RMS is lower than climatology over the tropical Pacific (Fig. 8g) and higher south of
 370 Australia and New Zealand (Fig. 8h). As the synoptic disturbances evolve, these initial anomalies
 371 contribute to the later-stage pattern: weaker variability over southern North America (Fig. 8e) and
 372 stronger variability over the extratropical South Pacific (Fig. 8f), respectively. Together, the ENSO
 373 composites suggest that changes in the spatial pattern of initial disturbance activity may influence
 374 downstream synoptic variability, particularly the Pacific meridional dipole patterns.



375

376 **Figure 8.** (a) The RMS anomalies (relative to climatology) of 200 hPa geopotential perturbations
 377 (shading; non-dimensional) of the first 50 optimal modes during their evolution ($t = 0-9$ days)

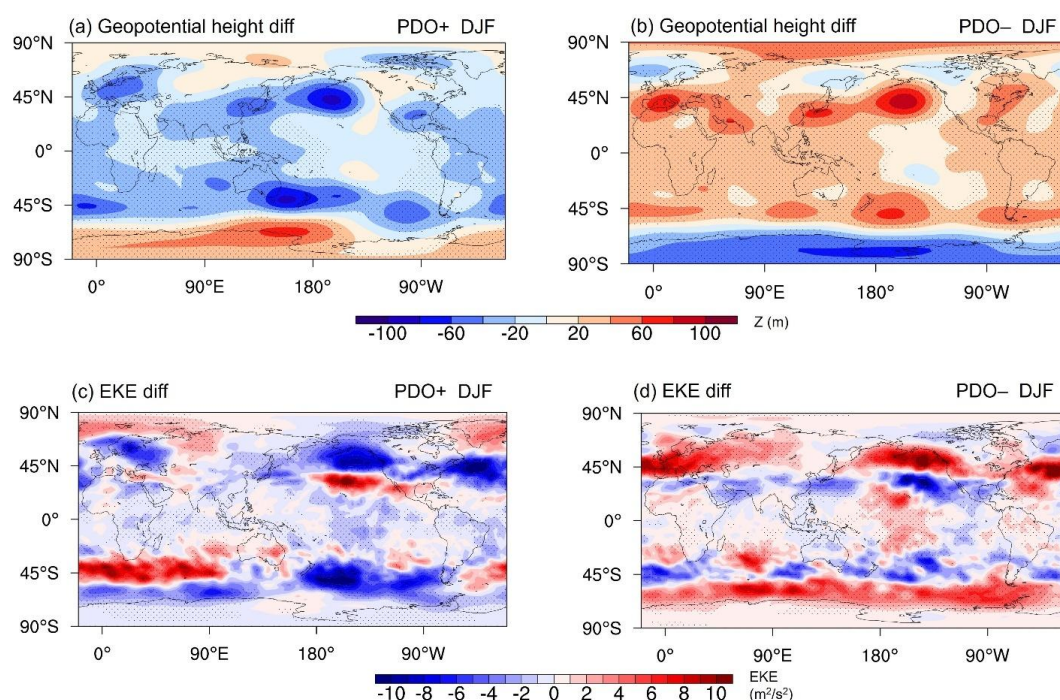


under El Niño-year background flow during DJF in the Northern Hemisphere. Stippling indicates regions where RMS anomalies are significant at the 5% level based on the two-tailed Student's *t*-test. Contours represent composite synoptic EKE (unit: $\text{m}^2 \text{s}^{-2}$) anomalies (relative to climatology) at 200 hPa during DJF for El Niño years in reanalysis. The contour ranges from -10 to 10 (unit: $\text{m}^2 \text{s}^{-2}$). **(b)** As in **(a)**, but for the Southern Hemisphere. **(c–d)** As in **(a–b)**, but for the initial stage ($t = 0–1$ day). Note that panels **c–d** use a different color bar from panels **a–b**. **(e–h)** As in **(a–d)**, but for optimal modes obtained from the La Niña year background flow.

385

We also examine decadal variability associated with the PDO (see Section 2.1). During positive PDO (PDO+) years, geopotential height anomalies are negative in northern and southern mid-latitudes and positive over Antarctica (Fig. 9a), while the negative phase (PDO–) broadly reverses this pattern (Fig. 9b). During PDO+, relative to the climatology, EKE decreases over the mid-latitude North Pacific and North Atlantic, with positive anomalies just south of the North Pacific minimum; in the Southern Hemisphere, EKE decreases over the South Pacific (Fig. 9c). During PDO–, northern mid-latitude EKE anomalies generally reverse, and a strong anomalous band appears over the Southern Ocean (Fig. 9d).

394

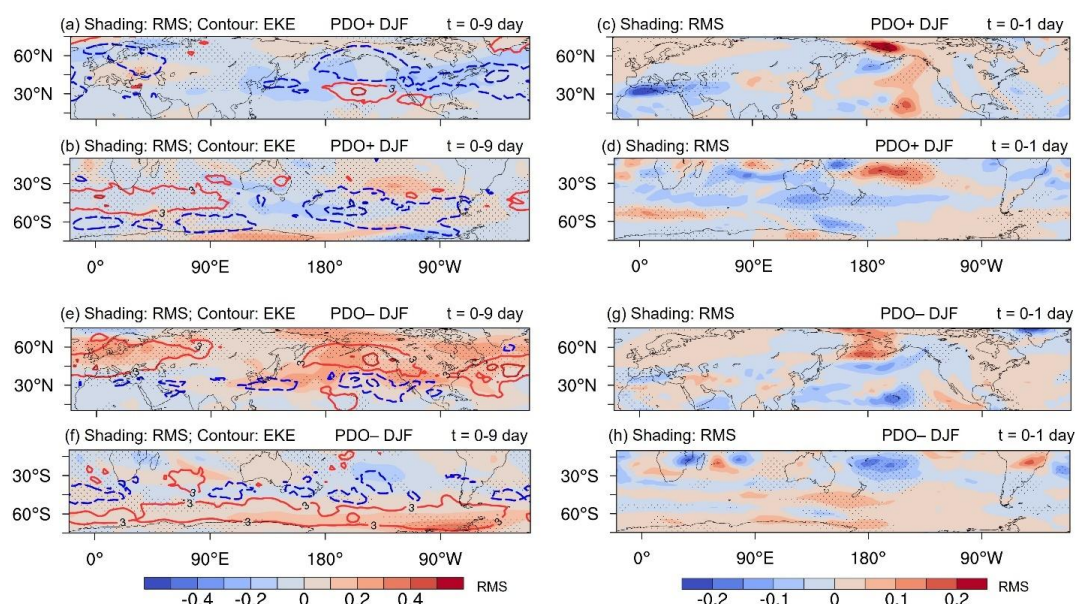




395 **Figure 9. (a–d)** As in Fig. 7, but for composite anomalies during the PDO+ and PDO– years.

396

397 For PDO+, 10-day RMS anomalies are mainly negative across East Asia, the Pacific, and
 398 North America (shading in Fig. 10a), consistent with observed EKE (contours in Fig. 10a). The
 399 optimal analysis does not reproduce the negative EKE anomalies over Europe or positive
 400 anomalies over the central North Pacific. In the Southern Hemisphere, it captures part of the
 401 negative South Pacific EKE pattern, but its center is displaced westward; it also misses the
 402 observed meridional contrast over the Indian and Southern Oceans (Fig. 10b). Thus, the PDO+
 403 modes only partly capture the observed spatial structure of EKE anomalies. These discrepancies
 404 may partly arise because the observed PDO+ EKE anomalies reflect disturbances and nonlinear
 405 interactions beyond those represented by the leading optimal modes. During the initial stage, RMS
 406 increases over Alaska and the eastern North Pacific but decreases near the Aleutians (Fig. 10c).
 407 The initial negative RMS anomalies (Fig. 10c) may contribute to the subsequent negative RMS
 408 anomalies across the Pacific–North America sector (Fig. 10a). Likewise, negative initial anomalies
 409 south of Australia and New Zealand (Fig. 10d) may contribute to the negative RMS anomalies
 410 during a 10-day period over the South Pacific (Fig. 10b).



411

412 **Figure 10. (a–h)** As in Fig. 8, but for the PDO+ and PDO– years.



413

414 During PDO–, 10-day RMS anomalies are positive across the northern mid-latitudes (shading
 415 in Fig. 10e), broadly consistent with the observed EKE anomalies (contours in Fig. 10e). At the
 416 initial stage, positive RMS anomalies occur over the Bering Sea (Fig. 10g). As the optimal modes
 417 evolve over the 10-day period, this initial enhancement may contribute to stronger synoptic
 418 variability farther downstream across the northern mid-latitudes. In the Southern Hemisphere, the
 419 10-day RMS captures the observed band of enhanced EKE over the Southern Ocean (Fig. 10f).
 420 Initial RMS anomalies are also positive south of Australia and New Zealand (Fig. 10h). Their
 421 subsequent development may likewise lead to stronger Southern Ocean variability, particularly
 422 near the Antarctic Peninsula (Fig. 10f).

423 **4. Conclusions**

424 We use optimal mode analysis to identify regions favorable to the initial growth of synoptic
 425 disturbances under climatological and climate-mode-related background flows. The modes
 426 reproduce key synoptic characteristics, and their RMS amplitude broadly captures the spatial
 427 pattern of synoptic variability. Under Northern Hemisphere climatological flow, three leading
 428 modes are initially concentrated over tropical–subtropical Asia and propagate toward the North
 429 Pacific and North America; their initial region shifts northward and growth weakens in JJA
 430 compared to DJF. In the Southern Hemisphere, DJF modes originate over the South Atlantic–
 431 southern Africa sector or tropical South Pacific, whereas JJA modes originate over the tropical
 432 Indian Ocean and Maritime Continent and move toward the South Pacific and South America.

433 The first 50 optimal modes sample a broader set of growing synoptic-scale perturbations.
 434 Their 10-day RMS broadly reproduces the principal storm tracks: in the Northern Hemisphere, it
 435 is strongest over the North Pacific and eastern North America–North Atlantic sector in DJF and
 436 weaker in JJA. In the Southern Hemisphere, DJF RMS lacks a pronounced maximum but is
 437 relatively high over the mid-latitude Indian Ocean; JJA RMS peaks over the subtropical South
 438 Pacific. Initial-stage RMS highlights regions favorable for disturbance development, typically
 439 located equatorward of their subsequent amplitude maxima. In the Northern Hemisphere, the
 440 initial regions are concentrated over tropical–subtropical Asia, with stronger growth in DJF than
 441 in JJA. In the Southern Hemisphere, they shift from the South Atlantic–southern Africa sector in
 442 DJF to the tropical Indian Ocean–Maritime Continent in JJA.



Initial-stage RMS also differs among ENSO and PDO phases. During El Niño, enhanced RMS over the tropical eastern Pacific and reduced RMS over the extratropical North Pacific coincide with stronger observed EKE over southern North America and weaker EKE farther north, respectively. An initial South Pacific meridional dipole corresponds to downstream variability there. During La Niña, reduced initial synoptic variability over the tropical Pacific may contribute to weaker downstream variability over southern North America, whereas enhanced initial variability south of Australia and New Zealand may contribute to stronger downstream variability over the extratropical South Pacific. Contrasting PDO-related initial anomalies near the Aleutians and south of Australia and New Zealand may contribute to subsequent variability across the northern mid-latitudes and the South Pacific–Southern Ocean, respectively.

5. Discussion

This study shows how background flow shapes regions favorable to initial synoptic disturbance growth and their subsequent pathways across seasons and climate states. Several tropical and subtropical regions favor initial perturbation growth before downstream developments. Agreement between optimal-mode RMS and observed EKE varies by season, hemisphere, and climate mode. Previous studies have mainly diagnosed the energy, wave activity, and propagation of identified disturbances (e.g., Pinheiro et al., 2021; Okajima et al., 2022; Ndarana et al., 2023; Zhuge et al., 2023; Battalio and Lora, 2024; Lee and Nakamura, 2024; Senapati et al., 2024; Yang and Li, 2025; Xue et al., 2026). The current framework based on optimal mode analysis complements these diagnostics by locating favorable initial growth and assessing perturbation amplification under a prescribed background flow. Compared to nonlinear optimization approaches such as conditional nonlinear optimal perturbation (Mu et al., 2003; Mu and Duan, 2005), it is less computationally demanding and yields multiple leading modes that enable a systematic assessment of preferred disturbance initiation regions and their downstream evolution across seasons and climate states.

The current framework has several methodological limitations that motivate further investigation. First, the QG formulation is best suited for extratropical disturbances and does not adequately represent equatorial or cross-equatorial dynamics. Because the QG approximation becomes less applicable toward the equator, the low-latitude preferred initial regions identified here should be interpreted as regions favorable for the subsequent growth of mid-latitude synoptic



473 disturbances within the QG framework, rather than as a full representation of tropical disturbance
474 dynamics. Future work using barotropic or primitive-equation frameworks applicable to tropical
475 dynamics would help assess the robustness of these low-latitude source regions. Second, the
476 numerical formulation uses second-order centered finite differences. Grid-convergence tests and,
477 where feasible, comparisons with spectral discretization would strengthen confidence in the
478 inferred disturbance growth and pathways. Third, the optimal mode depends on methodological
479 choices, including the growth interval, norm, linearization, and prescribed background flow
480 (Farrell, 1989). Our sensitivity tests cover the number of selected modes, pressure levels, damping,
481 and a 1–5-day optimization period, but do not examine alternative norms. Future work should
482 systematically evaluate these choices to quantify the robustness of the identified preferred initial
483 regions and subsequent evolution.

484 Several limitations related to the physical interpretation and application of the results also
485 warrant further investigation. First, the ENSO and PDO comparisons characterize disturbance
486 behavior under their associated composite background flows but do not isolate the effects of SST
487 forcing. Moreover, the relatively small composite samples and overlap among climate-mode
488 phases may limit attribution of regional differences to individual climate modes. Targeted
489 numerical experiments with controlled SST forcing and larger samples would help separate the
490 dynamical effects of individual climate modes. Second, although the present analysis identifies
491 preferred initial regions and their subsequent spatial evolution, the physical mechanisms linking
492 these regions to downstream synoptic variability can be further analyzed through energetics
493 diagnostics. Future work could track individual optimal modes in more details and analyze their
494 energy budgets, building upon Mak and Zhao (2022). Third, the current framework includes only
495 dry dynamics, but latent heating can substantially affect disturbance growth and structure (You
496 and Deng, 2025; Auestad et al., 2026). Comparing moist and dry optimal modes under identical
497 background flows would clarify whether latent heating modifies the preferred initial regions and
498 subsequent disturbance evolution. Finally, the preferred initial regions identified here represent
499 dynamically favorable regions for perturbation growth, providing a potential implication for
500 designing initial-condition perturbations in ensemble forecasts. Perturbations introduced in these
501 regions may undergo stronger amplification and subsequently influence downstream synoptic
502 variability, potentially improving the representation of forecast uncertainty. However, whether



503 such targeted perturbations can improve ensemble spread and forecast skill remains to be tested in
504 future studies.

505

506 **Code and data availability.** All the data used in this study are freely available online. The ERA5
507 reanalysis data is from <https://doi.org/10.24381/cds.6860a573>. The ONI index is from
508 https://www.cpc.ncep.noaa.gov/products/analysis_monitoring/enso/oni/v6/. The PDO index is
509 from <https://www.ncei.noaa.gov/pub/data/cmb/ersst/v5/v6/index/ersst.v6.pdo.dat>. The codes have
510 been uploaded in <https://doi.org/10.5281/zenodo.23262264>.

511

512 **Author contributions.** SZ designed and performed the analyses, and prepared the manuscript. QZ,
513 HA and JZ revised the manuscript and provided helpful comments.

514

515 **Competing interests.** The authors have no competing interests to declare.

516

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