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# **Underlying Physical Mechanisms of Winter Precipitation Extremes over India's High Mountain Region**

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1 **Abstract**

2 Extreme precipitation events (EPEs) are among the most pervasive weather hazards in the western  
3 Himalayan region (WHR), posing widespread damage to life, infrastructure, and agriculture. This  
4 study investigates the synoptic and large-scale characteristics linked to winter precipitation  
5 extremes over the WHR. EPEs are identified as events surpassing the 95<sup>th</sup> percentile threshold. A  
6 composite analysis is employed using two reanalyses—ERA5 and IMDAA to elucidate the  
7 synoptic conditions conducive to EPEs. Our findings suggest that EPEs in the WHR are linked to  
8 an intensified subtropical westerly jet, characteristically shifted to south than normal. Enhanced  
9 kinetic energy in the upper troposphere, attributed to increased baroclinic instability, reinforces  
10 moisture convergence and strengthens synoptic scale circulation, triggering deep convection and  
11 supporting EPEs. Notably, the interplay of pronounced Rossby waves sinking over the region,  
12 coupled with regional orography, significantly modulates the intensity of western disturbances  
13 (WDs) during extremes. Employing clustering analysis, we observed that the strongest EPEs are  
14 linked to anomalous vorticity in the upper to middle troposphere, together with deep convection  
15 via highly strengthened WDs, suggesting a potential role of large-scale influences. Using  
16 Lagrangian method, we identify that Arabian Sea is primary moisture source for EPEs in WHR.  
17 We further delved into the role of large-scale connections and EPEs through quasi-resonant  
18 amplification (QRA) analysis in the WHR using ERA5 data. The findings unveil the association  
19 of QRA with notably magnified, quasi stationary mid-latitude planetary waves characterized by  
20 wavenumbers 6/7/8 (baroclinic waves), contributing to precipitation extremes. Remarkably,  
21 distinct fingerprints of meridional temperature gradients, indicative of QRA, are linked to EPEs.  
22 Furthermore, this investigation discerns distinctive QRA patterns associated with varying clusters  
23 of extreme event intensities. Overall, our results emphasize the crucial role of QRA in amplifying  
24 planetary waves and promoting extreme precipitation in the WHR, underscoring the vulnerability  
25 of the region to evolving climate conditions and providing insights into the underlying physical  
26 mechanisms.

27 **Keywords:** Extreme precipitation events, western disturbances, western Himalayas, baroclinic  
28 instability, Rossby waves, Quasi-resonant amplification

29

30

## 31 **1. Introduction**

32 The Western Himalayan region (WHR, Figure 1a) significantly influences the hydro-  
33 meteorological conditions and climate variability in north India and adjoining regions. The diverse  
34 and topographic landscape of the WHR encompasses forests, cultivated areas, wetlands, glaciers,  
35 and urbanized zones (Figure 1b). This further affects the land-atmospheric exchange processes and  
36 precipitation variability (e.g. Singh et al. 1995; Beniston 2003; Anders et al. 2006; Dimri 2012;  
37 Hunt et al. 2018a; Nischal et al. 2022). In winter (December to February; DJF), western  
38 disturbances (WDs) –extratropical synoptic weather systems, contribute to a significant amount of  
39 annual precipitation to the region (e.g. Hunt et al. 2018a; Nischal et al. 2022). Moving eastward  
40 along upper tropospheric sub-tropical westerlies, these synoptic-scale eddies gather moisture from  
41 the Arabian Sea and encounter the regional orography across the WHR (e.g. Madhura et al. 2015;  
42 Hunt et al. 2018a). This intensifies the WDs and results in heavy precipitation (Ramaswamy 1956).

43 Winter precipitation in the WHR sustains the crucial glacial mass equilibrium, influencing  
44 regional river discharge and streamflow (Hasson et al. 2014). Given the vulnerability to  
45 precipitation variability, any key fluctuations could severely impact regional freshwater  
46 availability downstream in this vital glacier-dependent watershed, in turn, affecting millions (e.g.  
47 Messerli et al. 2004). Moreover, changing climate hold profound implications for this ecosystem  
48 and winter precipitation patterns (Tewari et al. 2017; Hunt et al. 2020), including the anticipated  
49 increase in hydroclimatic variability, modified precipitation patterns and intensified extremes  
50 (Miller et al. 2012; Madhura et al. 2015; Krishnan et al. 2019). Additionally, the steep topography  
51 of the WHR increases its susceptibility to intense surface runoff during extreme precipitation  
52 events (EPEs), enhancing the risk of avalanches, landslides, and floods, thus, rendering the WHR  
53 as a high-risk zone (Bookhagen and Burbank 2010; Priya et al. 2016; Acharya et al. 2023).  
54 Numerous instances of EPEs over the WHR have illustrated massive losses through cloudbursts  
55 induced by terrain-locked deep convective systems in valleys, and flash floods triggered by  
56 extratropical disturbances (Dimri et al. 2017; Houze et al. 2017; Hunt et al. 2021).

57 Recent studies suggest a rise in the frequency and intensity of winter EPEs in the WHR  
58 (Madhura et al. 2015, Shekhar et al. 2017, Krishnan et al. 2019, Rao et al. 2021; Nischal et al.  
59 2023). Some attributable factors include a warming climate (Ballesteros-Cánovas et al. 2018),  
60 growing prevalence of atmospheric rivers (Nayak et al. 2021) and, increased WD variability  
61 through enhanced upper-level baroclinicity (Madhura et al. 2015). However, empirical evidence

62 is insufficient to firmly establish a trend in WD frequency, suggesting potential changes in  
63 intensity. The disastrous impacts of EPEs extend to both natural and anthropogenic ecosystems,  
64 resulting in damage to life, infrastructure, crops, and power networks, sparking serious concerns  
65 about their impacts on human lives (Dimri et al. 2021; Sati and Kumar, 2022). For instance, Figure  
66 1c (Disastrous Weather Events catalogue by India Meteorological Department) illustrates the  
67 human mortality rate linked to winter extreme snowfall events over the WHR since the 1980s.  
68 Goklany (2009) suggested improvement in monitoring, forecasting, and preparedness for these  
69 events to potentially lower mortality rates, as evident in recent years (see Figure 1c). Furthermore,  
70 a holistic understanding of the associated dynamics and thermodynamics of EPEs can improve the  
71 accuracy of early warning systems which eventually helps in effective risk mitigation. Winter  
72 precipitation over the WHR is modulated to a large extent by various localized, synoptic and large-  
73 scale meteorological processes including variations in the sub-tropical westerly jet and WD  
74 activity, as well as more localized features such as cloudbursts. While the precipitation extremes  
75 and their associated dynamics have been relatively well-documented for the summer monsoon  
76 season (Priya et al. 2016; Revadekar et al. 2016; Vellore et al. 2016; Hunt et al. 2018b; Aggarwal  
77 et al. 2022), winter precipitation extremes and their causal mechanisms have received  
78 comparatively less attention (Madhura et al. 2015; Hunt et al. 2018b; Krishnan et al., 2019).  
79 Additionally, most available literature focuses on case studies, highlighting location or event-  
80 specific EPEs (e.g. Norris et al. 2015), underscoring the importance of in-depth research to address  
81 the potential ramifications of winter EPEs.

82         Recent research has also emphasized the critical role of large-scale planetary atmospheric  
83 dynamics in characterizing such weather extremes. Studies conducted by Petoukhov et al. (2013),  
84 Coumou et al. (2014), and Mann et al. (2018) shed light on Quasi-Resonant Amplification (QRA),  
85 a phenomenon linked to climate change associated Arctic warming i.e. Arctic amplification (AA).  
86 QRA occurs when these quasi-stationary atmospheric waves become trapped in a latitudinal  
87 waveguide (turning points around 30°N and 45°N), creating a zonally directed waveguide for  
88 specific wave numbers. When the waveguide is (almost) circumglobal, wave energy is efficiently  
89 trapped and waves constructively interfere with the forcing, leading to resonance and the growth  
90 of trapped planetary waves that are excited by thermal or orographic forcing (Petoukhov et al.  
91 2013). This phenomenon contributes to weakening/meandering of mid-latitude westerlies,  
92 enhancing the possibility of weather extremes. Francis and Vavrus (2012) found that rapid Arctic  
93 warming weakens the poleward thickness gradient, decelerating large-scale Rossby waves. This,

94 in conjunction with enhanced wave amplitudes and elongated meridional flow, contributes to more  
95 frequent atmospheric blocking patterns and mid-latitude weather extremes. Zonal wavenumbers  
96 6-8 have been known to resonate and amplify during summer extremes (e.g. Petoukhov et al. 2013;  
97 Mann et al. 2018). However, the influence of QRA on winter extremes over WHR is not  
98 investigated yet. Additionally, it has been observed that the strength of AA is relatively stronger  
99 during winter (e.g. Cohen et al. 2014), thus emphasizing the possible influence of QRA  
100 occurrences.

101 In this work, we investigate the synoptic and large-scale characteristics, including  
102 dynamical, thermodynamic, and moisture convective processes, associated with winter  
103 precipitation extremes in the WHR using high-resolution climate datasets. Collaterally, the study  
104 also evaluates the potential of recently released high-resolution Indian Monsoon Data Assimilation  
105 and Analysis (IMDAA) reanalysis in capturing these extremes over the WHR as well as their  
106 underlying mechanisms. At the end, we analyze the QRA characteristics to understand the  
107 potential links between planetary scale waves (induced by AA) and winter EPEs. This sheds light  
108 on how high-latitude dynamics influence EPEs in the WHR. Such understanding of precipitation  
109 extremes over high mountain region and their associated physical mechanisms is crucial for  
110 interpreting climate-change scenarios of extremes.

## 111 **2. Data and Methods**

### 112 **2.1. Data**

113 The study of precipitation extremes over the WHR requires high-resolution datasets, as the  
114 complex and heterogeneous Himalayan orography results in substantial spatial variability of both  
115 mean and extreme precipitation (Andermann et al., 2011). Here, we conducted an analysis of  
116 winter (DJF) EPEs using multiple high-resolution gridded datasets from different sources over the  
117 WHR (27.5-37.5°N and 72.5-80.5°E, see Figure 1a), from 1979 to 2019 except for Integrated  
118 MultisatellitE Retrievals (V3) for Global Precipitation Measurement (GPM-IMERG) dataset  
119 which spans 2000 onwards. To obtain daily precipitation observations, we used the India  
120 Meteorological Department's (IMD) dataset, which is based on measurements from 6955 rain  
121 gauge stations spread throughout the Indian subcontinent, interpolated to a resolution of  
122 0.25°×0.25° (Pai et al. 2014). Nevertheless, the density of stations in the WHR is low. We also  
123 utilized the GPM-IMERG dataset, a high-resolution merged satellite product that combines  
124 microwave and infrared observations from the GPM satellite constellation with gauge observations

125 using the Day-1 IMERG algorithm (Huffman et al. 2015). IMERG is available from 2000 onwards  
126 at half-hourly temporal and  $0.1^\circ$  spatial resolution. The satellite product has been incorporated, as  
127 previous studies have evaluated its performance for studying precipitation extremes over complex  
128 topographies (Liu et al. 2020; Nepal et al. 2021).

129 Additionally, we used a recently released regional reanalysis dataset, IMDAA, which has  
130 a high spatial resolution of 12 km and covers the South Asian domain. IMDAA was developed  
131 through a collaboration between the National Centre for Medium Range Weather Forecasting, the  
132 UK Met Office, and IMD, using a unified atmospheric model and the four-dimensional variational  
133 (4D-Var) data assimilation technique (Rani et al. 2021). The dataset provides better representation  
134 of orographic features due to its high resolution (Nischal et al. 2022). Lastly, we employed the  
135 state-of-the-art global reanalysis ERA5, developed by the European Centre for Medium Range  
136 Weather Forecasts (fifth generation), which has a resolution of  $0.25^\circ \times 0.25^\circ$ . We also considered  
137 daily values of various meteorological variables from IMDAA and ERA5, such as air temperature,  
138 specific humidity, vorticity, and three-dimensional wind components, at different pressure levels.  
139 Additionally, ERA5 and IMDAA-based WD tracks from Nischal et al. (2022) were utilized to  
140 identify and filter out the WD centers during EPEs over the WHR. Daily averages of outgoing  
141 longwave radiation (OLR) data from KALPANA-1 satellite ( $0.25^\circ \times 0.25^\circ$ ) for 2004-2019 have  
142 also been used.

## 143 **2.2. Methodology**

### 144 **2.2.1. Composite analysis of extremes**

145 A wide discrepancy in precipitation patterns is observed among different datasets over the WHR  
146 (Baudouin et al. 2020; Nischal et al. 2022). Thus, we focus on analyzing how different datasets  
147 depict precipitation extremes over the WHR. Considering that the selected datasets are generated  
148 with different input data and dissimilar developmental methods, a strong agreement between any  
149 two of them indicates they are likely to be close to reality (Baudouin et al. 2020; Nischal et al.  
150 2022). We also explore the fidelity of the newly-developed high-resolution IMDAA reanalysis in  
151 representing WH precipitation extremes during the winter season. It is to be noted that high  
152 resolution (spatial and temporal) as well as comparatively long temporal coverage provides the  
153 possibility of relatively better depiction of precipitation extremes (Rani et al. 2021; Nischal et al.  
154 2022). Additionally, its ability to simulate both small and large-scale atmospheric dynamics (as it



155 represents the complex interplay between topography and mountain meteorology) during winter  
156 (Nischal et al. 2022) and summer (Saini and Attada, 2023) precipitation is credible.

157 We have selected datasets from different sources with at least daily precipitation records  
158 and high spatial resolution ( $\leq 25$  km) for the analysis. Extreme days are defined as those when the  
159 average regional (27.5°N-37.5°N and 72.5°E-80.5°E) precipitation exceeds the 95<sup>th</sup> percentile  
160 threshold (Fig. 2a). This threshold is calculated by considering the precipitation values (including  
161 zero precipitation) across all the grid points in the region during the investigation period.  
162 Conversely, we classify the remaining days as non-extreme days. We further examined the  
163 composites of different atmospheric variables for extreme minus non-extreme days in the ERA5  
164 and IMDAA reanalysis, using IMD-identified (observation-based) days in ERA5 and IMDAA's  
165 own identified days. Due to its shorter time span, IMERG was exclusively utilized for analyzing  
166 precipitation characteristics and not for further composite analysis. Nonetheless, it is noteworthy  
167 that the identified extreme and non-extreme days in IMERG exhibit similarities to other datasets  
168 for the period 2000-2019.

### 169 2.1.2. *k*-means clustering algorithm

170 In this study, we also seek to identify distinct weather regimes associated with extreme  
171 precipitation over the WHR, for which we use *k*-means clustering. Neal et al. (2020) previously  
172 used a similar technique to identify different weather patterns over India. The *k*-means clustering  
173 algorithm (Hartigan and Wong, 1979) is an iterative unsupervised vector quantification algorithm  
174 that groups a given set of *n*-dimensional (vector) points into distinct non-overlapping clusters (*k*),  
175 based on nearest possible cluster centroid value. The algorithm uses the Euclidean distances of  
176 each point in the cluster to measure similarities between them. The cluster points are grouped such  
177 that the sum of square of the distances for each point to the cluster mean or centroid gets  
178 minimized. The objective function, *F*, that the *k*-means clustering algorithm seeks to minimize is  
179 defined thus:

$$180 \quad F = \sum_{j=1}^k \sum_{i=1}^{n_j} \|x_{i,j} - \mu_j\|^2 \quad (1)$$

181 where  $x_{i,j}$  is a point vector in cluster *j* and  $\mu_j$  is the center of cluster *j*, having  $n_j$  points. Firstly, the  
182 algorithm initializes the *k*-cluster centers randomly and each *x* point vector is grouped to a certain  
183 cluster *j* based on the closest cluster centroid value. Following this, the centroid values are

184 recalculated according to the means of all assigned points in each cluster and the algorithm keeps  
185 iterating until convergence.

186 Three clusters were deemed to be most appropriate for this analysis, enough to explain a  
187 sufficient amount of the variance associated with precipitation extremes over the region (following  
188 Hunt et al. 2018a and heuristic ‘elbow method’ developed by Hardy, 1994). The clustering has  
189 been carried out over the region (72.5-80.5°E, 27.5-37.5°N) for the following variables:  
190 Geopotential Height (GPH), vertical velocity, Potential Vorticity (PV), divergence, and cloud  
191 cover fraction.

### 192 *2.1.3. Lagrangian parcel tracking for Moisture sources*

193 To identify possible moisture pathways associated with an EPE, we employed a Lagrangian parcel  
194 tracking approach developed by Hunt et al. (2018b). This involves decomposing the atmosphere  
195 above the event of interest into uniformly spaced parcels, which are subsequently backward-  
196 advected from the region of interest using ERA5 wind data. Following the determination of the  
197 start date, time, and location of interest, we extract the three-dimensional wind vector at this point.  
198 The parcels undergo back-advection with 20-minute integration time, by changing the signs of  
199 wind vector components using a set of spherical equations (see more details in Hunt et al. 2018b).  
200 Due to limited reanalysis resolution, the resulting point may lack precise data coordinates in space  
201 or time. To mitigate this, local reanalysis fields are recalculated for the correct time through cubic  
202 spline interpolation. Subsequently, we employ a tricubic method (Lekien and Marsden 2005) to  
203 estimate the wind vector at the desired location. This iterative process continues, returning to the  
204 initial step with updated time, location, and winds, persisting for the specified integration duration.  
205 Here, we have back-advected forty parcels, equally spaced between 925 and 500 hPa for 10 days  
206 using local, contemporaneous ERA-5 winds at hourly temporal scale.

## 207 **3. Results and Discussion**

### 208 **3.1. Precipitation extremes**

209 Figure 2(a) depicts the percentile distributions for daily precipitation exceeding the 80<sup>th</sup> percentile  
210 in different datasets over the WHR. Each line on the figure corresponds to a distribution that  
211 encompasses all the grid points across the study region (approximately 1850 grids). To ensure a  
212 fair comparison, the datasets have been re-gridded to a common spatial resolution of 0.25°. The  
213 findings indicate that a substantial proportion of winter precipitation in the WHR is sourced from

214 events surpassing the 80<sup>th</sup> percentile. Despite some differences, precipitation distribution shows a  
 215 consistent pattern across all datasets. IMERG tends to underestimate comparatively lower  
 216 percentiles but performs better than ERA5 at higher percentiles. IMD and IMDAA are often quite  
 217 similar to each other. Excessive precipitation, surpassing the 95th percentile, in such hilly terrains  
 218 can contribute to increased regional runoff, potentially causing downstream floods that can  
 219 subsequently impact the Rabi crops sown during winter (e.g. Haritashya et al. 2006; Arora et al.  
 220 2016). Thus, 95<sup>th</sup> percentile has been chosen as the threshold for categorizing precipitation  
 221 extremes and non-extremes.

222 Figure 2b illustrates how the composited region-wide precipitation anomalies during  
 223 extremes evolve over time in the WHR. The notable anomalies are detectable from day -6 to day  
 224 -2, with a more pronounced and remarkable increase occurring from day-2 to day 0. Following  
 225 day 0, the anomalies decline sharply until day +2. The entire life cycle of these EPEs generally  
 226 lasts roughly 4-5 days; similar to lifecycle of WDs (see Dimri et al. 2016). The evolution structure  
 227 appears to exhibit a slight asymmetry between the advancing and decaying stages, suggesting that  
 228 comparatively more robust characteristics and mechanisms are evident during the growing phase  
 229 (e.g. Xu et al. 2022). The advancing phase also appears to be slightly slower than the decay phase,  
 230 which aligns with the observed asymmetric precipitation footprints associated with a WD. This  
 231 suggests the significant impact of WD characteristics on the winter precipitation in the WHR. A  
 232 substantial rise in precipitation amounts during extremes is observed over the region as  
 233 demonstrated by different datasets. WDs, suggested to be primary baroclinic cyclonic storms,  
 234 develop and intensify through the release of energy via the atmospheric baroclinic response and  
 235 their interaction with WHR orography (Hunt et al. 2018a), potentially contributing to precipitation  
 236 extremes. Therefore, we also explored the temporal evolution of the composited region-wide  
 237 anomalies for baroclinic instability using the baroclinic instability criterion (C), based on vertical  
 238 wind shear and potential temperature (Phillips, 1954) in the upper (200 hPa) and lower troposphere  
 239 (700 hPa), following Madhura et al. (2015). The baroclinic instability criterion (C) is defined thus:

$$240 \quad C = \frac{f^2(u_{200}-u_{700})\theta_{500}}{\beta gH(\theta_{200}-\theta_{700})} \quad (2)$$

241 Here, ‘f’ denotes the Coriolis parameter, ‘ $\theta$ ’ is potential temperature, ‘u’ refers to the zonal wind  
 242 speed, ‘ $\beta$ ’ is the meridional gradient of Coriolis parameter, ‘g’ is the acceleration due to gravity  
 243 and H is vertical height (m) between the 200 and 700 hPa pressure levels.

244           The evolution structure demonstrates a rise of baroclinic instability in the region four days  
245 preceding the occurrence of extreme precipitation. This enhancement becomes notably  
246 pronounced two days prior to the event and reaches its peak intensity one day prior. As the extreme  
247 precipitation day approaches, there is a notable and very sharp decrease in the baroclinic instability  
248 in the region, which continues to diminish until the following day. These findings indicate that  
249 enhanced baroclinicity serves as a leading indicator for the occurrence of such extremes. This  
250 increase in baroclinic response of the atmosphere can contribute to intensification of WDs and  
251 result in enhanced precipitation, provided sufficient atmospheric moisture is available. The  
252 growing baroclinicity can possibly influence the vertical component of relative vorticity and cause  
253 intense convection in the moist middle and upper troposphere, ultimately contributing to  
254 precipitation extremes (e.g. Para et al. 2019).

255           Higher positive anomalies for precipitation amounts are observed in the geographical  
256 distribution of composited precipitation anomalies on day 0 for all four datasets (Figure 2c-f). The  
257 highest differences are evident along the WHR foothills, where the climatological precipitation  
258 maxima are situated (Nischal et al. 2022); however, diversity in precipitation patterns among  
259 different datasets is quite evident. Regarding the geographical distribution, IMDAA and ERA5  
260 reanalyses exhibit close agreement, indicating heavy precipitation across the entire orographic  
261 band- specifically lower Himalayas and foothills. However, IMD places heavy precipitation more  
262 towards the north and northwest, while IMERG indicates it in the foothills and lower elevations.  
263 It is noteworthy that IMD encounters challenges due to the lack of weather stations, leading to  
264 estimates derived from downslope extrapolation (Kishore et al. 2016). Satellite-based microwave  
265 retrievals, on the other hand, often faces difficulties in accurately assessing precipitation over  
266 snow-covered areas and estimating cold season orographic precipitation (e.g. Derin et al. 2016).  
267 Generally, differences in data generation sources and development algorithms create such  
268 variabilities among datasets (Baudouin et al. 2020; Nischal et al. 2022). Nonetheless, it is clear  
269 that precipitation extremes induce widely distributed heavy precipitation over the region, and it is  
270 crucial to understand their physical mechanisms.

### 271 **3.2. Possible Mechanisms**

272           To better understand the physical processes for precipitation extremes in the WHR, a diagnosis of  
273 synoptic and large-scale characteristics was performed. Composite anomalies for extreme minus  
274 non-extreme days were analyzed to gain insight into the associated physical conditions.

### 275 **3.2.1. Synoptic Characteristics**

#### 276 **a) Geopotential height, potential vorticity and winds**

277 GPH anomalies are distinctive signatures of deep synoptic cyclonic troughs (Cannon et al., 2014;  
278 2015), also known as WDs. Increased variability in upper and middle tropospheric GPH is  
279 associated with increased winter WD activity and precipitation (e.g. Lang and Barros, 2004;  
280 Cannon et al., 2015). Here, we present the composite maps of upper (200 hPa) and mid-  
281 tropospheric (500 hPa) GPH anomalies for precipitation extremes minus non-extremes in ERA5  
282 and IMDAA (Figure 3a-d). At both levels, there is a marked negative GPH anomaly, with a  
283 minimum situated just west of the WHR at roughly 34°N-68°E (200hPa). The troughs formed are  
284 considerably more intense at 200 hPa than in the mid-troposphere. IMDAA and ERA5 exhibit a  
285 consistence in trough placement. The locations of WD centers (20°–40°N, 60°–82°E) during EPEs  
286 in ERA5 and IMDAA reveals that the majority of these centers are situated west of the WHR,  
287 aligning with the trough placements in respective datasets (Figure 3a-b). Notably, both ERA5 and  
288 IMDAA exhibit a concentrated pattern for WD centers, although minor disparities can be observed  
289 in terms of the precise locations of these centers within the datasets.

290 Positive PV in the upper-level corresponds to cyclonic circulation, while negative values  
291 usually suggest anticyclonic circulation. Moreover, a prominent meridional gradient of upper-level  
292 PV can strengthen the background flow, facilitating the expansion of cyclonic circulations to lower  
293 levels too (Hoskins 1997; Hoskins et al. 2007; Attada et al. 2022). The upper-tropospheric  
294 (300hPa) PV composite anomalies, depicted in Figure 3(e-f), illustrate the presence of strong  
295 positive PV anomalies along the WHR, providing conditions that support moist convection. The  
296 reduced atmospheric stability through higher PV flux have been linked to initiation of enhanced  
297 convection through Rossby wave-breaking (Attada et al. 2022). In particular, the interaction of  
298 WDs and stronger PV can help in the growth of WDs through moist baroclinic instability via  
299 orographic interactions during extremes.

300 Further, we examined meridionally-averaged (27.5-37.5°N), zonally distributed vertical  
301 structures of composited GPH anomalies (Figure 4a-b). A robust pressure trough, slightly tilted  
302 eastwards, is evident across the western Himalayan longitudes, with a core at centered at about  
303 300 hPa. Such strong negative GPH anomalies are indicative of intense synoptic trough formation  
304 and the strengthening of potent cyclonic disturbances, and the structure is consistent with that of  
305 strong WDs (Hunt et al. 2018a). The interaction between the vertically-tilted deep trough and

306 regional topography induces orographic lifting of the flow, further intensifying the WDs through  
307 moist baroclinic instability. This process leads to heavy precipitation over the region, provided  
308 sufficient moisture supply is available (Lang and Barros, 2004; Cannon et al. 2014; Baudouin et  
309 al. 2021). Furthermore, we analyzed the differences in vertical wind speed between extreme and  
310 non-extreme days using a vertical-longitudinal cross section (Figure 4c). A stronger ascent  
311 indicating increased deep convection during extremes is observed over the WHR, with a maximum  
312 at about 450 hPa, located over steep orography, underscoring the crucial contribution of orographic  
313 forcing to the moist flow. In general, advancing WDs are associated with a pronounced ascent at  
314 the forefront, followed by a large-scale descent in the rear (Hunt et al., 2018a). During EPEs, these  
315 ascent and descent patterns become more intense, underpinning the essential role played by intense  
316 WDs during such events.

317 As WDs are carried to the WHR along the subtropical jet, the latitudinal position of the jet  
318 in turn affects the location and interactivity of WDs with orography, thus affecting precipitation  
319 patterns and intensity over the region (Krishnan et al. 2019). Thus, we examined the latitude-  
320 pressure cross-sections of zonally averaged (72.5°-80.5°E) zonal wind (200 hPa) composites  
321 during extremes and non-extremes (Figure 4d-e). Both ERA5 and IMDAA composites show that  
322 during extremes, the subtropical jet intensifies and shifts further south from its mean position over  
323 the WHR in winter. This affects WD activity through enhanced atmospheric baroclinicity and  
324 reduced stability, favoring the development of precipitation extremes.

### 325 **b) Cloud cover and outgoing longwave radiation (OLR)**

326 Dimri (2013) reported that subzero temperatures on snow-covered surfaces during winter hinder  
327 convective activity necessary for deep cloud activity over the WHR. Consequently, migrating  
328 WDs account for the majority of cloud cover in the region (Hatwar et al. 2005; Madhura et al.  
329 2015; Sankar et al. 2021). Figure 5a-d displays composite anomalies based on ERA5 and IMDAA  
330 data, revealing changes in TCC and OLR across the WHR, respectively. During extremes, TCC  
331 experiences a widespread increase of up to 48%, indicating intensified precipitation.  
332 Simultaneously, OLR values exhibit a regionwide decrease, peaking in the zone of maximum  
333 observed precipitation, indicating the presence of deeper convection and higher cloud tops. Similar  
334 patterns are evident for OLR in satellite dataset KALPANA-1 (Figure S1). As OLR exhibits a  
335 clear sky dependence on surface feedback during winter, these negative anomalies signify the  
336 influence of stronger WDs contributing to increased precipitation (Dimri 2013).

337 **c) Transient eddy kinetic energy and baroclinic instability**

338 The variability of the large-scale circulation over the WHR and surrounding regions is  
339 characterized by higher-frequency transient eddies (i.e. WDs), which largely grow through the  
340 conversion of available potential energy into kinetic energy via baroclinic instability (Pedlosky,  
341 1972; Chang and Orlanski, 1993). We investigated the variations in 200 hPa eddy kinetic energy  
342 (EKE) in IMDAA and ERA5 to highlight the localized impacts of winter westerly wave activity  
343 and subsequent energy transformation/exchange processes in the atmosphere. EKE is commonly  
344 described as the kinetic energy associated with the time-varying component of the horizontal  
345 velocity field.

$$346 \quad EKE = \frac{1}{2}(u'^2 + v'^2) \quad (3)$$

$$347 \quad u = \underline{u} + u' \text{ and } v = \underline{v} + v'$$

348 where,  $u$  and  $v$  are horizontal velocity components,  $u'$  and  $v'$  denote time-varying velocity  
349 components whereas,  $\underline{u}$  and  $\underline{v}$  represent the time mean velocity components. As our focus in this  
350 study is towards understanding the role of kinetic energy associated with synoptic transient eddies  
351 in driving precipitation extremes over the WHR, a Lanczos filter of 2-10 days has been applied to  
352 the anomalies of horizontal velocity fields. Composite anomalies for EKE (Figure 6a-b) indicate  
353 the accumulation and availability of regions with positive and negative kinetic energy in the  
354 atmosphere, indicating a Rossby wave train pattern. Increased EKE over the WHR indicates the  
355 existence of strong WDs, as we have already seen. This increase in EKE provides evidence for the  
356 presence of intense WDs over the region. The interplay of these WDs with strong PV gradients  
357 potentially fuels their growth through moist baroclinic instability. This, in turn, intensifies  
358 convection and EKE, ultimately resulting in substantial contributions to heavy precipitation.

359 It is well established that the subtropical jet is associated with strong upper-atmospheric  
360 baroclinicity and a large meridional temperature gradient, which help in the growth and  
361 maintenance of WDs (Singh and Agnihotri, 1977). A baroclinically unstable environment over the  
362 WHR favors a further intensification of these WDs and can lead to heavy precipitation (Hunt et al.  
363 2018a,b; Sankar et al. 2021; Rao et al. 2022). Here, we explored the baroclinic nature of the  
364 atmosphere during EPEs using the baroclinic instability criterion (C) described earlier (Equation  
365 1). Figure 6(c-d) depicts the differences in mean baroclinic instability between composites of  
366 precipitation extremes and non-extremes. The anomalies depict a baroclinic wave train, with a

367 pronounced indication of baroclinic instability over the WHR. Baroclinic instability has been  
 368 proposed as a potential mechanism for release of energy from available potential energy in the  
 369 atmosphere, thereby aiding in the growth and intensification of WDs (e.g. Rao and Rao, 1971).  
 370 Hence, the presence of positive anomalies over the WHR implies an increased baroclinic activity,  
 371 which indicates a stronger WD activity in the area, leading to heavy precipitation (e.g. Madhura et  
 372 al. 2015). Given that the wave train extends far - beyond the WHR in both ERA5 and IMDAA,  
 373 this indicates potential links to large-scale phenomenon that might contribute to a consistent  
 374 intensification of WDs during precipitation extremes.

#### 375 **d) Moisture Transport Dynamics**

376 Baroclinic instability is the primary driver of WDs, but their intensity might not always correlate  
 377 with the intensity of observed precipitation levels, as it is contingent on moisture availability. Some  
 378 winter disturbances with strong winds and temperature fluctuations may result in less intense  
 379 precipitation, while slower-moving, less intense baroclinic lows may carry moisture over long  
 380 distances, causing more prolonged precipitation. Catastrophic EPEs sometimes involves a long  
 381 preconditioning process, accumulating moisture advected from distant sources within a subsident  
 382 non-precipitating environment (e.g. Turato et al. 2004). The moisture then gets released in a  
 383 concentrated area, such as the WHR, with moisture flux convergence arising out of a sudden  
 384 synoptic forcing consequent to baroclinic development and the concomitant role of local  
 385 orography. Winter precipitation over the WHR is primarily contributed through moisture  
 386 advection from the Arabian Sea with secondary contributions from the Mediterranean, Caspian,  
 387 and Red Seas (Dimri et al., 2015; Barlow et al. 2005). Here, we investigate moisture supply to the  
 388 WHR during precipitation extremes through the examination of composite anomalies for vertically  
 389 integrated moisture transport (VIMT;  $\text{kg m}^{-1}\text{s}^{-1}$ ).

$$390 \quad VIMT = \frac{1}{g} \int_{surface}^{300} qV dP \quad (4)$$

391 where,  $q$  is specific humidity,  $V$  is the horizontal wind, and  $dp$  is the vertical incremental change  
 392 in pressure.

393 The composited anomalies in moisture transport, computed from surface to 300 hPa in  
 394 ERA5 and IMDAA reanalysis, utilize surface pressure as the lower boundary, offering a more  
 395 robust approach, particularly in the context of the complex topography of the WHR. Our analysis  
 396 highlights the Arabian Sea to be a major moisture supply source to these extremes, which is



397 consistent with the findings of Hunt et al. (2018b) using back trajectory analysis (Figure 7a-b).  
398 Furthermore, a correlation analysis of these daily VIMT anomalies with the frequency of  
399 precipitation extremes over the WHR exhibits a strong correlation for moisture transport from the  
400 Arabian Sea (Figure 7c-d). Notably, both ERA5 and IMDAA produced almost identical results for  
401 VIMT and correlation analysis, indicating our findings are robust. Examining lead/lag composites  
402 of VIMT anomalies unveils a moisture flow from distant westward sources up to 4-5 days prior  
403 the event. Subsequently, moisture contributions become more pronounced from relatively nearby  
404 sources, especially the Arabian Sea, as the event day approaches (Figure S3). This suggests that  
405 an increase in the synoptic variability of WD associated westerly moisture transport on daily  
406 timescales is an essential precursor of precipitation extremes. The dynamics of moisture  
407 contributions for different intensity extreme events is further discussed in section 3.2.2.

408 Further, we utilize a Lagrangian approach (discussed in section 2.1.3) to investigate the  
409 moisture sources for three intense extreme precipitation events observed over the WHR. The  
410 selected cases under examination occurred at the following locations and dates: (a) 32.25°N,  
411 76.5°E, 12 Dec 2017; (b) 32.75°N, 75.75°E, 5 Feb 2013; and (c) 33.25°N, 74.75°E, 5 Dec 2006.  
412 For each case study, we computed forty parcel trajectories backward for 10 days using local,  
413 contemporaneous ERA-5 winds (one-hourly data) and examined specific humidity along the  
414 trajectories (Figure 8). The findings unveil that air parcels are dispersed across Europe and North  
415 Africa, extending into neighbouring moisture sources, ten days preceding the event.  
416 Predominantly, these parcels originate near the North Sea and the Mediterranean Sea and are  
417 subsequently transported as boundary layer parcels distributed over the Black Sea, Caspian Sea,  
418 and the Arabian Gulf. Notably, there is substantial mid-/lower-tropospheric convergence  
419 associated with the events, indicative of the passage of a WD. Examination of specific humidity  
420 profiles along these trajectories highlights a significant moisture contribution from distant sources  
421 such as the North Sea (part of North Atlantic Ocean), Mediterranean Sea, and Black Sea. However,  
422 a noteworthy proportion of moisture is observed to originate specifically from the Arabian Sea.  
423 Overall, it can be suggested that the temporal scales of moisture accumulation may vary  
424 significantly depending on the scale of the precipitating event, with sources of moisture ranging  
425 from more to less remote, leading to corresponding variations in the temporal scales for moisture  
426 advection.

### 427 **3.2.2. Classification of Extremes: k-means clustering**

428 An important aspect of our study is to attempt to categorize winter precipitation extremes over the  
429 WHR. Such classification can provide useful insights into the different types of driving dynamics  
430 for extremes, specifically the underlying synoptic and large-scale mechanisms. To achieve this,  
431 we have applied k-means clustering algorithm to various dynamic parameters over extreme days,  
432 following Hunt et al. (2018a). The parameters incorporated in our analysis are: GPH, vertical  
433 velocity, PV, divergence, and cloud cover fraction (72.5-80.5°E, 27.5-37.5°N). The first step is to  
434 construct a standardized distribution for each field by computing standardized anomalies at each  
435 grid point over the study region for the entire multi-level (1000-100 hPa) time series of the extreme  
436 precipitation days, since the given fields have different statistical distributions. The anomaly fields  
437 have been re-gridded vertically to produce voxels with equal volumes to maintain homogeneity  
438 and prevent overpopulating lower altitudes with higher level density. Before clustering, the multi-  
439 dimensional array values in the field are unraveled and then concatenated into a single vector.  
440 Cluster mean vectors are then obtained by running the algorithm on this single vector and we then  
441 reconstruct the original three-dimensional fields from these. Here, we have classified the  
442 precipitation extremes into three clusters.

443 The vertical structures of various dynamic field clusters constructed using ERA5 are  
444 presented in Figure 9. The columns separate different clusters, arranged in increasing order of GPH  
445 anomaly in the atmosphere. Type 1, 2 and 3 clusters are separated by intensity in all the clusters  
446 where type 1 shows the weakest intensity extremes and type 3 shows strongest types of extremes,  
447 though being least common among all clusters. Type 2 which shows the extremes associated with  
448 relatively intermediate intensity, compared to other two clusters, seem to be the most common  
449 class of extremes over the WHR. Crucially, all clusters are somewhat dynamically similar,  
450 implying the presence of a common dynamical source of precipitation (i.e. WDs).

451 The zonally distributed vertical structure of GPH anomalies in the type 1 cluster shows a  
452 very weak, though vertically tilted trough over the region of interest, with a minimum over 66°E.  
453 A deepening of the trough aligned with a slight eastward shift in the location of the depression  
454 minimum can be observed as we move from type 1 to type 3 clusters. The depressions in type 2  
455 and 3 clusters are identical to those associated with the WDs over the region (Dimri and Chevuturi  
456 2016; Hunt et al. 2018a). The differences in their magnitudes indicate the strength of the WD in  
457 type 3 extremes being associated with the much intense WDs than type 2. Vertical wind in all three

458 clusters shows ascent over the region followed by a descent to the west, a pattern that intensifies  
459 as we move from type 1 to type 3.

460 A prominent upper-level maximum for PV over the study region can be observed in type  
461 2 (situated at about 200–250 hPa), and type 3 clusters (located at about 400–450 hPa), whereas it  
462 is almost non-existent in type 1. These structures in the upper troposphere have characteristics  
463 resembling those of baroclinic instabilities elsewhere in the atmosphere, and of WDs themselves  
464 (Molinari et al. 1995; Robinson, 1989; Hunt et al. 2018a). The maximum in PV along the elevated  
465 topography of the WHR in turn leads to orographic forcing (as we see in the vertical velocity  
466 composites), thus supporting intense precipitation over the region. The zonally distributed  
467 structures of divergence in the type 2 and 3 clusters are also baroclinic, as concluded earlier. The  
468 ascending and descent patterns observed earlier are favored with upper-level divergence (~200–  
469 250 hPa) ahead and convergence behind, identical to what we observe for a typical WD. These  
470 convergence patterns and pronounced ascent, on interaction with the regional orography, ensure  
471 favorable environment for triggering precipitation extremes. The patterns of cloud cover fractions  
472 confirm this inference as we observe highly intensified cloud cover over the WHR in type 2 and 3  
473 clusters, whereas type 1 cluster again shows relatively weaker patterns.

474 Further, we have tried to understand how these dynamical classifications relate to the  
475 observed precipitation patterns during extremes over the region. The patterns for mean  
476 precipitation during the extreme days for each cluster has been presented in Figure 9(p-r). As  
477 expected, type 1 cluster is associated with small-scale precipitation with weaker intensity, mostly  
478 reaching up to 12-24 mm/day. Type 3 cluster provides evidence of widely distributed large-scale  
479 precipitation with strong intensity (24-36 mm/day), whereas type 2 shows an intermediate kind of  
480 response. Overall, it becomes clear that the type 2 and type 3 clusters show strong evidence of  
481 association with WD structures and may be related to intense and very intense WDs. Lead-lag  
482 composites of VIMT anomalies have been constructed to elucidate moisture contributions from  
483 their sources for various event types. The results reveal that type 1 events exhibit a relatively  
484 greater moisture flow from distant westward sources preceding the events, in contrast to type 2  
485 and 3 events (Figure S4-S6). Additionally, more intense events display a notably higher moisture  
486 supply from relatively nearby sources, particularly the Arabian Sea, as the event day approaches.  
487 The substantial variations in precipitation magnitudes between type 2 and 3 clusters suggest the  
488 existence of additional factors, beyond the presence of robust WDs, that potentially contribute to

489 the enhancing the intensity of these WDs. Therefore, our subsequent investigation will concentrate  
490 on comprehending the large-scale attributes linked to EPEs across the WHR.

### 491 **3.2.3. Large scale characteristics**

#### 492 **a) Response of Rossby wave source to precipitation extremes**

493 The role of large-scale circulation in producing precipitation extremes over north Indian region  
494 during winter has been suggested to be linked with co-existence of jet with an upper-tropospheric  
495 quasi-stationary Rossby wave train (e.g. Hunt et al. 2018b). During winter, upper-tropospheric  
496 convergence and strong vertical motion in the subtropics generate anomalous vorticity (Nie et al.  
497 2019). This upper-level vorticity source, denoted as Rossby wave source (RWS), sets off Rossby  
498 wave trains which circulate the tropical heating to extratropical atmospheric circulation  
499 (Sardeshmukh and Hoskins, 1988; Nie et al. 2019). Here, we have investigated the patterns of  
500 these RWSs during precipitation extremes over the WHR and surrounding regions. The RWS  
501 function is defined thus:

$$502 \quad \frac{\partial \zeta_a}{\partial t} - v_\psi \cdot \nabla \zeta_a = -\zeta_a D - v\chi \cdot \nabla \zeta_a \quad (5)$$

503 where,  $-\zeta_a D$  denotes the vortex stretching (relates to local strong divergence), the second term on  
504 the left-hand side is the advection of vorticity gradient by rotational wind (caused by large-scale  
505 divergent flow), which is related to Rossby wave propagation. The two right-hand side terms are  
506 forcing terms, together comprising the RWS.

507 The examination of RWS composite anomalies for precipitation extremes (Figure 10a)  
508 indicates that the WHR experiences strong negative anomalies in RWS, while adjacent regions  
509 show weak RWS, an amplification of the climatological patterns as observed by Nie et al. (2019)  
510 for winter RWS over the region. This suggests that strong WDs – themselves Rossby wave-like  
511 features – are dissipating over the WHR. During winter, Rossby wave sinks generated over the  
512 WHR are associated with negative anomalies for vortex stretching and positive anomalies for  
513 advection of vorticity gradient (e.g. Shimizu and de Albuquerque, 2010; Nie et al. 2019). This  
514 implies that large-scale divergent flow is the primary factor responsible for the generation of wave  
515 sinks in this region (see Ding et al. 2023). The interplay of anomalous PV flux, possibly from  
516 higher latitudes, with WDs in the vicinity of the WHR helps to strengthen the WDs associated  
517 local vortices. This contributes to intensified low-pressure systems and extreme precipitation.  
518 Therefore, the interaction of large-scale planetary circulation patterns and WDs over the

519 orographic regimes of the WHR influences the wintertime precipitation intensities in this region.  
520 Nonetheless, it is important to note that the western Himalayan topography has a significant  
521 influence on the variability patterns related to these Rossby wave trains.

#### 522 **b) Wave activity flux (WAF)**

523 Large-scale, quasi-stationary high amplitude anomalies in the atmosphere can contribute to  
524 abnormal weather patterns by affecting migratory weather systems and meridional flow  
525 exchanges. The release of energy as a stationary Rossby wave train by these anomalies can  
526 potentially result in the formation of atmospheric blocking patterns, with converging WAF often  
527 indicating the development of a blocking anticyclone. Thus, converging/diverging patterns in  
528 WAF linked to stationary Rossby waves on a meandering zonal mean flow can aid in  
529 comprehending the underlying dynamics governing the genesis of these large-scale circulation  
530 anomalies (Takaya and Nakamura, 1997). Additionally, the zonally asymmetric mean flow aids  
531 the growth of these anomalies by converting available potential energy to kinetic energy via  
532 baroclinic instability processes. The meridional eddy fluxes of momentum and heat play a key role  
533 in the maintenance and forcing of atmospheric flow, and the Eliassen-Palm (EP) flux can be used  
534 as a diagnostic tool to study their interaction. The composites of zonally averaged EP flux during  
535 extreme and non-extreme days have been examined (Fig. 10b-c). The propagation of Rossby wave  
536 groups, resembling the time-averaged life cycle of nonlinear baroclinic waves (Simmons and  
537 Hoskins, 1980; Hoskins, 1983), was observed to move upwards from the surface at lower levels  
538 and turn equatorward above 400 hPa. This pattern of EP cross sections has been found to be linked  
539 with northward PV flux at lower levels and southward PV flux with strong wind shear in the upper  
540 troposphere. Significantly, the flux is much stronger during extreme days as compared to non-  
541 extreme days, indicating a stronger baroclinicity during extreme events that facilitates the  
542 development and sustenance of cyclonic systems. Furthermore, a deceleration of zonal wind is  
543 observed over the mid-latitudinal troposphere, which is much stronger during extreme events (Fig.  
544 10d-e; contours).

545 Stronger EP convergence is also noted over almost the entire troposphere from 20° N  
546 towards the poles, with EP divergence on the poleward side at lower levels (Fig. 10d-e). This  
547 convergence is primarily related to energy flux and indicates that waves tend to slow down the  
548 zonal mean flow with zonal energy being transformed into wave energy (Smith et al. 2022). The  
549 deceleration is found to be much more pronounced during extreme events, indicating a higher

550 baroclinicity with more decelerated and/or meandering jet, and more energy available to fuel the  
551 persistence of cyclonic systems. The weakening/deceleration of mid-latitude westerlies has been  
552 linked to Arctic-sea loss due to Arctic warming, indicating an increased risk of extreme weather  
553 events (Smith et al. 2022). This generates decelerated quasi-stationary wave patterns that point to  
554 the existence of atmospheric blocking patterns characterized by a robust zonal flow both to the  
555 north and south of the blocking systems, and a shift from a zonal to a meridional flow pattern  
556 during initialization and decaying stages. During extreme days between 0-90°E and 20-80°N, a  
557 seemingly omega type of blocking pattern emerges, featuring a three-point vortex system (tripole)  
558 with an anticyclonic point vortex (high) situated on the poleward side of two cyclonic point  
559 vortices (Fig. 11a). This three-point vortex system can become stationary and guide the  
560 equatorward movement of cold polar air masses when the speeds become equal with the prevailing  
561 mid-litudinal westerly winds (Detring et al. 2020).

562         The upper-tropospheric (200 hPa) WAF introduced by Takaya and Nakamura (2001) can  
563 be employed to investigate the instantaneous propagation of both migratory and stationary quasi-  
564 geostrophic waves on a zonally asymmetric basic flow for precipitation extremes. This study  
565 focuses only on the horizontal components of the WAF (Fig. 11b), which is formulated using the  
566 stream function ( $\psi$ ) and the horizontal components of the basic flow wind vector ( $U$  and  $V$ ). A  
567 circumglobally propagating WAF appears to be strengthening from an anomalous RWS, linked to  
568 an omega blocking pattern situated over the North Pacific. The flux propagates towards the  
569 northeast, eventually passing over the central-eastern region of North America, the pattern being  
570 consistent with the Pacific/North American teleconnection (Berry 1995). Subsequently, the WAF  
571 splits, with one branch turning equatorward into the tropical Atlantic and other penetrating into  
572 northern latitudes towards Arctic region. The former branch of WAF follows a similar path to  
573 migrating WDs, growing stronger as it moves through Mediterranean region and propagates across  
574 Middle East, Iran, Iraq, Afghanistan, Pakistan, finally penetrating into the WHR. The latter branch  
575 weakens over the mid-Arctic Ocean but gains strength as it reaches the western parts of Eurasian  
576 continent, corresponding to the enhances stream function (Fig. 11b) and GPH anomalies (Fig. 11a),  
577 finally reaching the WHR. As demonstrated, EPEs over the WHR are notably influenced and  
578 amplified by both mid-latitude and higher latitude systems, implying the role of large-scale  
579 mechanisms.

### 580 c) Role of Quasi-resonant amplification

581 Atmospheric blocking patterns have been previously connected to longitudinally extended, slow  
582 or stationary quasi-stationary waves (QSW) in the atmosphere (Kautz et al. 2022). Quasi-resonant  
583 amplification (QRA), a specific type of QSW, is characterized by a double-peaked westerly jet  
584 profile with stronger westerlies in subtropical and subpolar regions and weaker ones in  
585 midlatitudes. A circumglobal jet profile with turning points at around 30°N and 45°N can  
586 efficiently guide and enhance waves with  $k=6/7/8$ , leading to resonance. When these trapped  
587 waves are excited by thermal or orographic forcing, they grow in amplitude. The waveguide  
588 condition depends on the zonal wave number  $k$  and the shape of the zonal mean zonal wind ( $\bar{u}$ )  
589 profile, linked to meridional temperature gradients in the lower troposphere through the thermal  
590 wind relationship (Petoukhov et al. 2013; Mann et al. 2018). Here, we investigated the previously  
591 unexplored association between QRA and winter EPEs in the WHR. Zonal wave number spectra  
592 for the meridional wind fields at 300 hPa over 30° to 45°N (Fig. 12a) depict the peak in amplitudes  
593 for wavenumbers 6-8 during extremes compared to climatology, indicating the favorable  
594 conditions for the QRA occurrence (Petoukhov et al. 2013, Coumou et al. 2014, and Mann et al.  
595 2018). Moreover, we explore the zonal wave zonal mean spectra for three types of EPEs previously  
596 classified via cluster analysis (Fig. 12b). Type 1 events, characterized by weak intensity  
597 precipitation, show amplitudes related to wavenumber 6 exclusively. Type 2 events, characterized  
598 by intermediate precipitation intensity, exhibit maximum amplitudes for wavenumbers 6, followed  
599 by 7 and 8 equally. Finally, type 3 events, with the most intense precipitation, display maximum  
600 amplitudes for wavenumber 8, followed by 7 and then 6, suggesting a highly meandering wave  
601 pattern during such extreme precipitation events. Additionally, the presence of a smaller-scale  
602 structure can be inferred from this pattern, which, in turn, can contribute to intensified quasi-  
603 geostrophic uplift due to the larger vorticity gradients observed.

604 Fig. 12(c-e) illustrates the interannual variability for the amplitudes of monthly mean  
605 Fourier components of 300 hPa meridional winds during individual winter months associated with  
606 wave numbers  $k=6,7$  and 8, averaged over the latitudinal range of 30-45°N using ERA5 data  
607 between 1979-2019. Linear regression (trend lines not shown in the plot) suggests moderate  
608 positive trend in most of the month-wise amplitudes in each category, except for January in wave  
609 numbers 7 and 8, which showed a slight negative trend. Based on 1.5 SD amplitudes, years  
610 depicting the possible QRA characteristics were identified. It is further noteworthy that nearly all  
611 the identified QRA years exhibit a notable increase in either intensity or frequency, or both, of

612 EPEs (Fig S2). These findings support the possible association between QRA phenomena and the  
613 magnitude or occurrence rate of such winter precipitation extremes over the WH. Further, the  
614 selected years were examined for QRA fingerprint behavior using 1000-hPa meridional  
615 temperature anomalies (Fig. 12f-h). In general, the QRA fingerprint pattern displays negative  
616 values in the subtropics, followed by a rise to neutral values around mid-latitudes, then a decline  
617 towards negative values through 50°N, and pronounced positive values again at higher subpolar  
618 latitudes (e.g. Mann et al 2018). The QRA fingerprints associated with wave number 6 months  
619 were slightly distorted for January and February but well depicted for December. For wave  
620 numbers 7 and 8, the identified years through 1.5SD amplitudes displayed a pronounced QRA  
621 fingerprint.

622 Further, we observed the double-peaked westerly jet profile using 300hPa zonally averaged  
623 (30-80°E) zonal winds for the selected years in each wave number type, in agreement with QRA  
624 characteristics, with stronger westerlies in subtropics as well as subpolar regions and, weaker  
625 westerlies in midlatitudes (Fig. 12i). Coumou et al. (2014) reported that the frequency of QRA-  
626 linked weather extremes has increased as a response to AA, which is driven by feedback  
627 mechanisms such as anthropogenic greenhouse warming and ice-albedo feedback. The amplified  
628 warming has a more profound effect on the polar boundary of the waveguide. Due to the thermal  
629 wind relationship between upper-level westerlies and lower tropospheric temperatures, this AA  
630 may reduce upper-level midlatitude westerlies ( $\bar{u}$ ), increasing the likelihood of QRA occurrences.  
631 The findings in our study suggest that QRA characteristics affect atmospheric dynamics and jet  
632 propagation, ultimately impacting winter precipitation over the WHR by altering Rossby wave  
633 characteristics.

### 634 **3.4 Exploring QRA associated EPEs: Case study**

635 Lastly, we investigated three instances of extreme precipitation over the last two decades, which  
636 persisted for at least three consecutive days and exceeded the 95<sup>th</sup> percentile threshold each day  
637 individually and given, that they fall into QRA years (Fig. 13). Additionally, we investigated if  
638 these events were linked to QRA characteristics. It is worth noting that the selected years align  
639 with those identified through the Fourier transform series based on 1.5 SD (standard deviation)  
640 amplitudes. The selected case studies include extreme precipitation occurrences on 2-5 December  
641 2006, 3-6 February 2013, and 10-12 December 2017. Figure 13 (a-i; shaded) shows the spatial  
642 distribution of precipitation anomalies before, during, and after the occurrence of EPEs over the



643 WHR. During the EPEs, widely distributed precipitation anomalies reaching up to 16 mm/day can  
644 be observed in certain regions, with little or no precipitation before and after the events. These  
645 heavy precipitation events are supported by presence of strong positive PV flux, which move away  
646 from the region as the event duration surpasses (Figure 13a-i; contours). The analysis of QRA  
647 characteristics associated with the events illustrates the presence of QRA fingerprint in all selected  
648 events, with the December 2017 events exhibiting the most pronounced fingerprint than the  
649 February 2013 event (Fig. 13j). This suggests that the meridional temperature profile has a  
650 substantial impact on the QRA fingerprint for these events, with Arctic-amplified warming  
651 projecting onto this latitudinal anomaly pattern. Moreover, the events were associated with a  
652 prominent wave 6/7/8 pattern in the zonal mean zonal wave spectra, with the December 2017 event  
653 having the strongest amplitudes and the February 2013 event having weaker amplitudes (Fig. 13k).  
654 The anomalies for 300hPa meridional winds also indicated the presence of wavenumber 6/7/8  
655 patterns, implying a connection to QRA characteristics (Fig. 13l-n). Lastly, the events also  
656 depicted the presence of a double-peaked westerly jet profile associated with QRA, with the  
657 February 2013 event having the least pronounced profile (Fig. 13o). Overall, the findings suggest  
658 that the observed EPEs over the WHR are associated with QRA characteristics, which may be  
659 influenced by Arctic amplified warming.

#### 660 **4. Summary and Conclusion**

661 Winter precipitation over the WHR is crucial for agricultural sustenance and recharging the  
662 western Himalayan glaciers. However, the projected rise in precipitation extremes with respect to  
663 climate change signal is a key matter of concern. Our study focuses on examining the associated  
664 synoptic and large-scale characteristics, including the impact of planetary scale dynamics, that  
665 contribute to extreme precipitation in the WHR. We aim to contribute to a much needed and deeper  
666 understanding of the fundamental physical mechanisms underlying these extremes. The key  
667 inferences from our study is as follows:

- 668 ● The majority of winter precipitation in the WHR is sourced from higher percentiles, with  
669 a substantial contribution from EPEs, amplifying the region's susceptibility to  
670 precipitation-related hazards.
- 671 ● During EPEs, the primary wave guide for WDs - the sub-tropical westerly jet - intensifies  
672 and shifts southward and helps in developing much stronger baroclinicity, consistent with

673 structures of strengthened WDs. This implies that synoptic dynamics play a key role in the  
674 increased variability of precipitation patterns over the WHR. The influence of local  
675 thermodynamics on precipitation intensification can be ascertained through the presence  
676 of higher kinetic energy and deeper convective activity during these events.

677 ● The analysis of potential moisture sources through the Lagrangian approach reveals that  
678 intense EPEs over the WHR are associated with diverse moisture sources, spanning from  
679 the North Sea and Mediterranean Sea to the Arabian Sea. Besides the established westward  
680 advection of moisture, our study underscores the pivotal role of the Arabian Sea as a  
681 significant moisture source for intense precipitation extremes in the WHR. The observed  
682 strong positive correlations between the frequency of precipitation extremes and moisture  
683 transport from the Arabian Sea highlight its crucial influence on such EPEs. Overall, the  
684 diverse moisture origins identified through our Lagrangian approach, emphasizes the  
685 intricate interplay of regional and distant moisture sources (with potentially varying  
686 temporal scales of transport) in shaping the dynamics of extreme precipitation over the  
687 WHR.

688 ● The classification of different types of precipitation extremes carried out using k-means  
689 clustering reveals that intense and very intense WDs are associated with heavy  
690 precipitation observed over the region, supported by high anomalous vorticity and deeper  
691 convection in the atmosphere.

692 ● Enhanced atmospheric instability during EPEs may be caused by the interplay of WDs with  
693 strong positive PV fluxes, possibly from higher latitudes, in the upper troposphere. This  
694 can help in further strengthening the WDs in these orographic regimes through moist  
695 baroclinic instability processes. The large-scale divergent flow resulting from these  
696 interactions can also potentially lead to the formation of Rossby wave sinks in the region,  
697 indicating the influence of large-scale planetary circulation patterns.

698 ● Large-scale, quasi-stationary anomalies can contribute to extreme precipitation in the  
699 WHR by decelerating the mean flow and generating blocking patterns that enhance the  
700 meridional flow exchanges. Such a phenomenon can contribute to further intensification  
701 of passing WDs over the WHR and lead to persistent extreme precipitation patterns.

- 702       ● Climate change associated arctic amplification may have a pronounced influence on the  
703 intensities of winter precipitation over the WHR through the QRA mechanism. Changes in  
704 meridional temperature gradients affect upper tropospheric wind profiles via the thermal-  
705 wind relationship. Extreme meandering QSW patterns in the wavenumber range 6/7/8  
706 demonstrate the role of planetary-scale dynamics in influencing precipitation intensity in  
707 the third pole's sub-regions (WHR), underscoring the influence of large-scale mechanisms.
- 708       ● The newly developed high-resolution reanalysis, IMDAA, realistically represents the  
709 regional precipitation distribution, trends as well as dynamical and thermodynamic  
710 contributions for precipitation extremes over the region.

711 In summary, our analysis highlights the importance of planetary, synoptic, and mesoscale  
712 processes as crucial drivers of extreme precipitation in the WHR. We observed that synoptic-scale  
713 dynamics become stronger during these events, which is further bolstered by local  
714 thermodynamics and large-scale Rossby wave dynamics. These factors work together to create  
715 conditions that promote the formation of intense weather systems with deeper convection and  
716 increased precipitation.

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## 726 **Conflict of interest**

727 The authors declare no conflict of interest.

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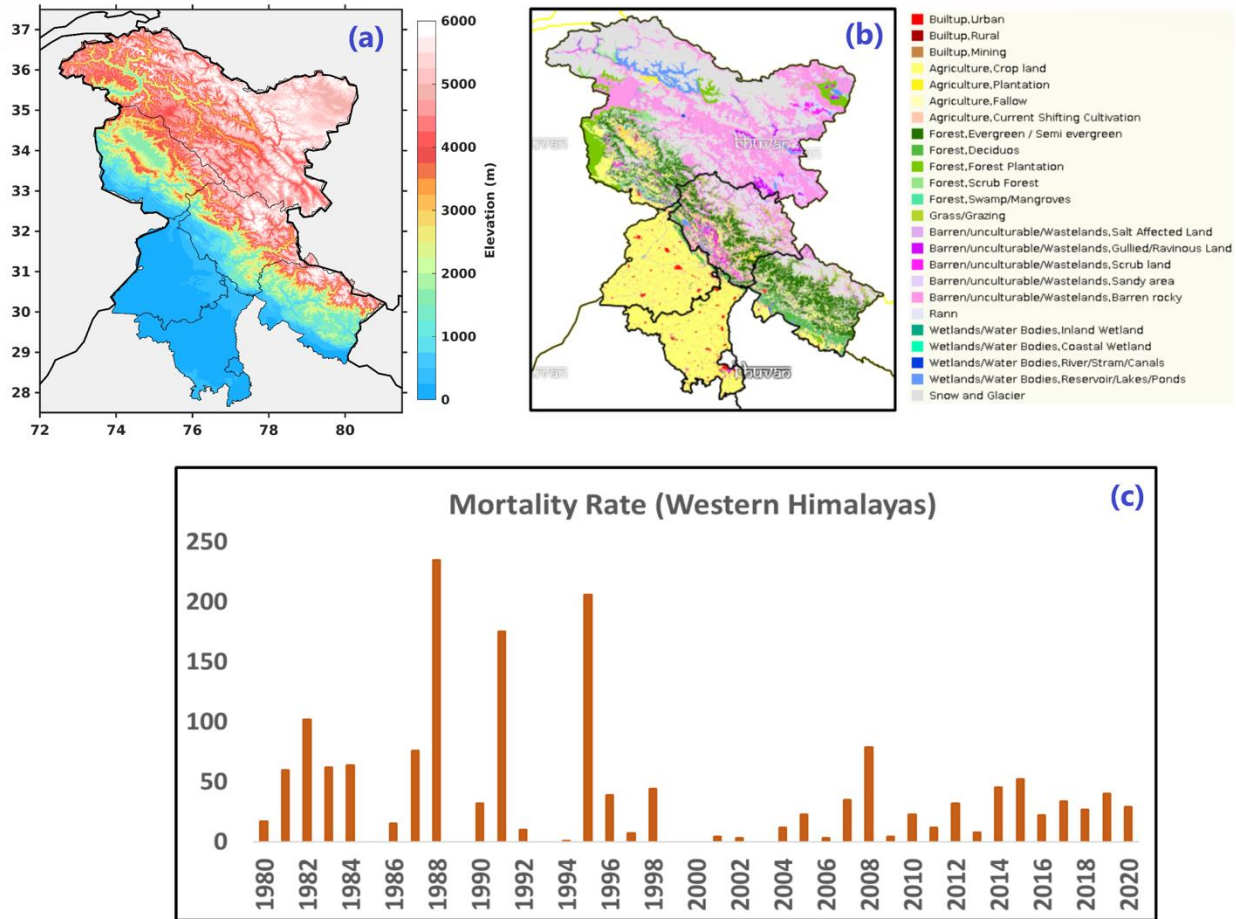
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986 **Figure 1:** Topographic map of the western Himalayan region (WHR; a), Land Use Land Cover map (b) for

987 the region from NRSC, Bhuvan (Indian Space Research Organisation). Figure 1c shows human mortality

988 rate due to extreme snowfall events over the WHR during the winter season between 1980-2020.

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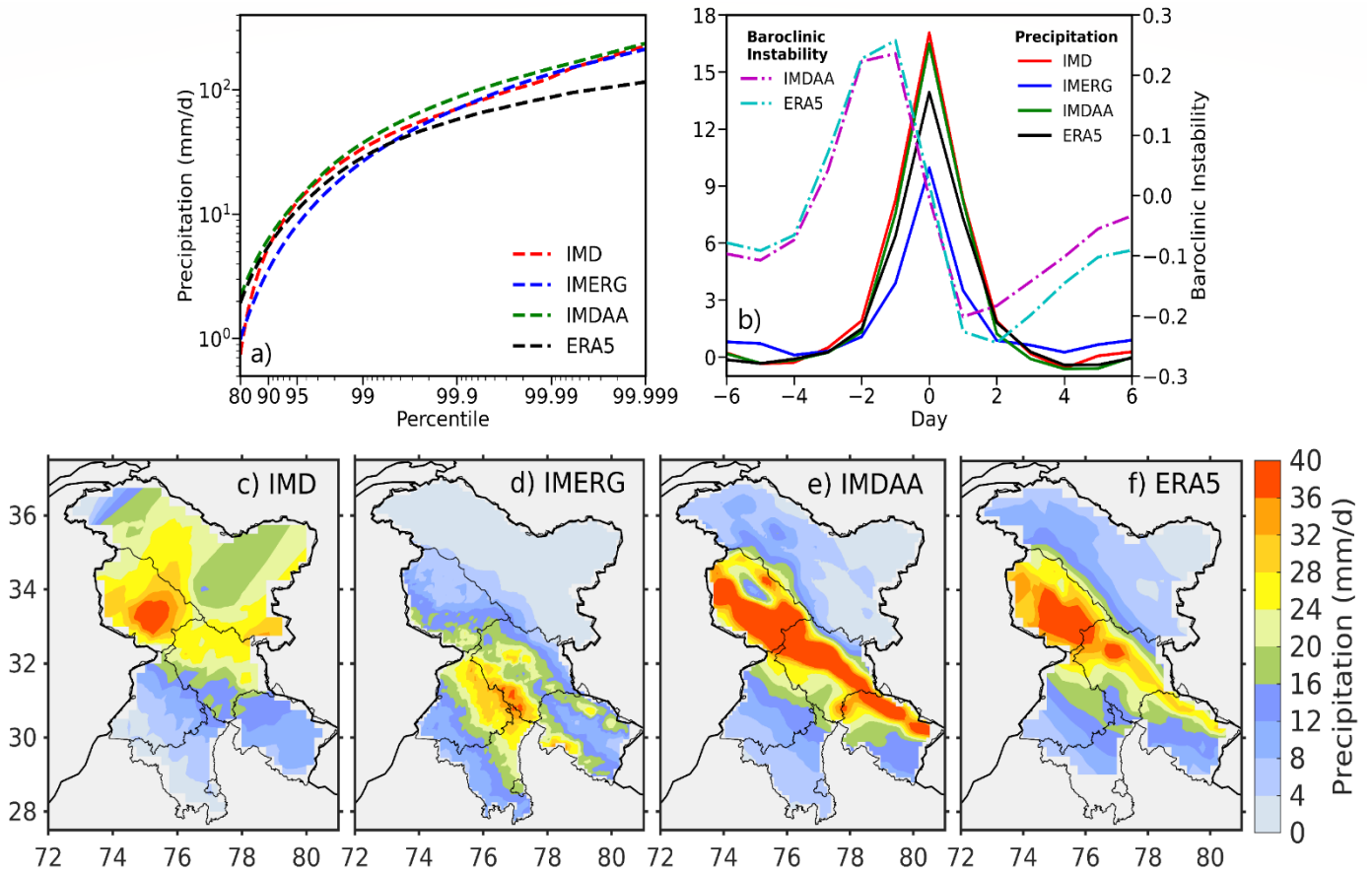
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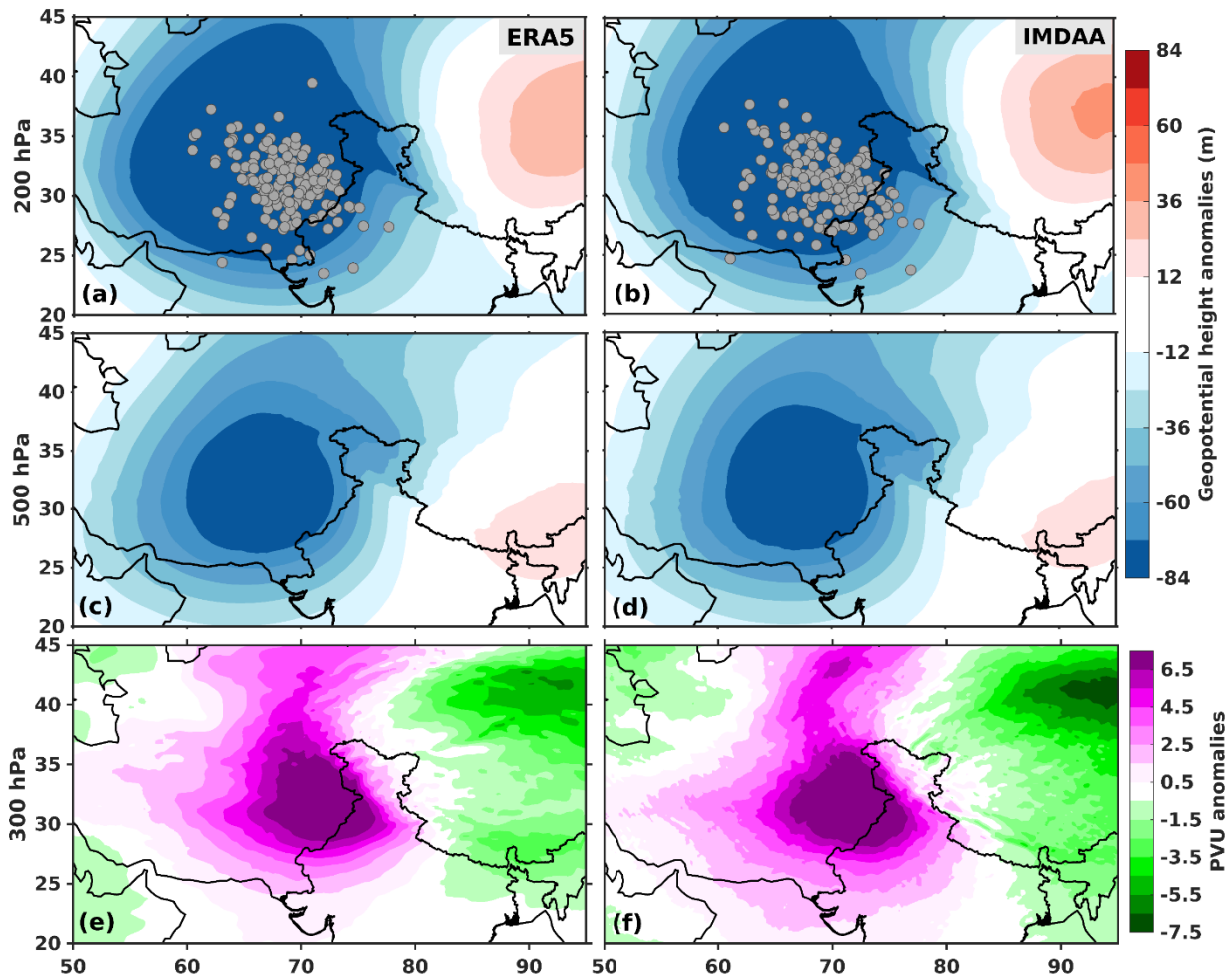
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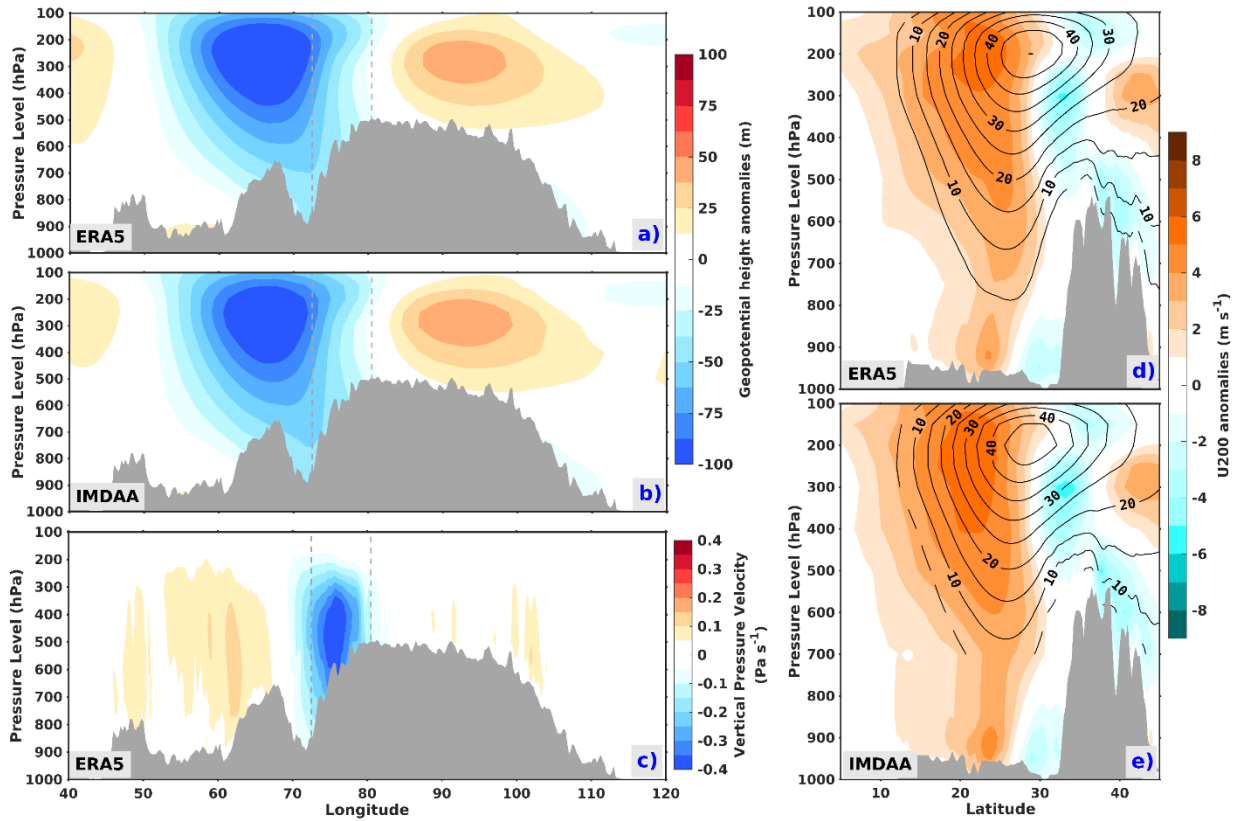
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**Figure 2:** Percentile distribution of winter (DJF) Precipitation (unit: mm/day) in IMD, IMDAA and ERA5 datasets during 1979-2019 and from 2001-2019 in GPM-IMERG over the WHR (a), daily scale evolution of regionally averaged precipitation anomalies during extreme precipitation days (b; solid lines). Dashed lines in (b) represent the regionally averaged baroclinic instability index anomalies during precipitation extremes. (c-f) Spatial distribution of composite precipitation anomalies (precipitation observed during extreme days (exceeding 95<sup>th</sup> percentile) minus non-extreme days) for winter precipitation.



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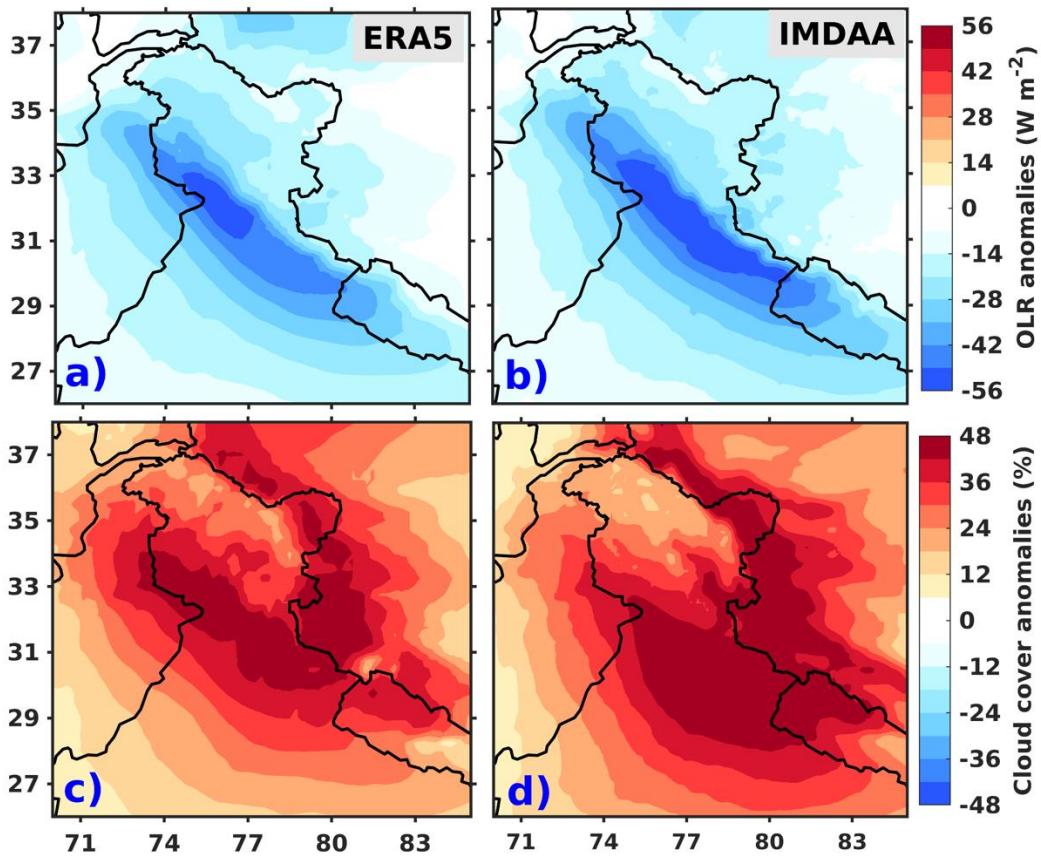
**Figure 3:** Composite anomalies (extremes minus non-extremes) for DJF geopotential height at 200 hPa (a-c, unit: m) and 500 hPa (b-d, unit: m) for the period 1979- 2019 and, 300 hPa baroclinic potential vorticity (e-f, unit: PVU). 1 PVU (potential vorticity units) here is equivalent to  $10^{-6} \text{ km}^2/\text{kg/s}$ . Western Disturbance centres during the extreme precipitation days has been represented using grey scatter points (a-b).



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**Figure 4:** Vertical structure of DJF geopotential height composite anomalies (a-b, unit: m) averaged latitudinally over 27.5°-37.5°N in ERA5 and IMDAA respectively and, vertical velocity (c, unit: Pa/s) in ERA5 from 1979-2019. The longitudinal bounds of the study region have been represented by dashed lines in grey (a-c). The panels (d) and (e) show the vertical structures for composite anomalies of zonal wind (U200) speed during extremes (shaded, unit: m/s) and non-extremes (contours, unit: m/s). Grey areas indicate the presence of orography.

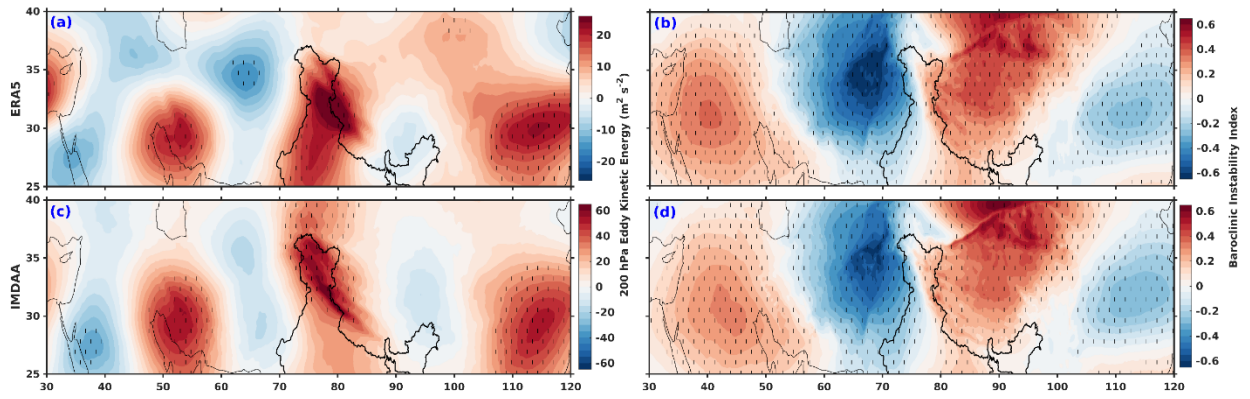




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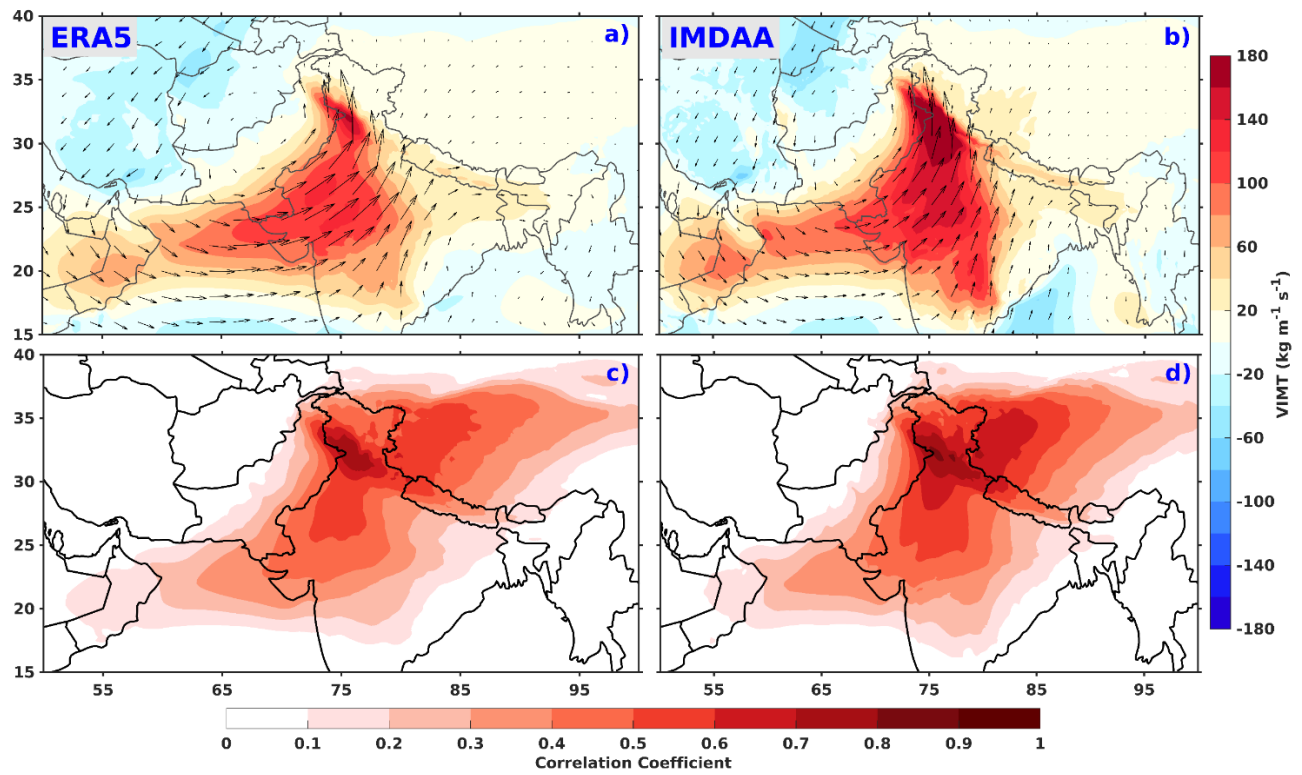
**Figure 5:** Composite anomalies (extremes minus non-extremes) for (a-b) total cloud cover (unit: %) and, (c-d) outgoing longwave radiation (OLR, unit:  $W/m^2$ ) in ERA5 and IMDAA datasets from 1979-2019 during winter season (DJF).





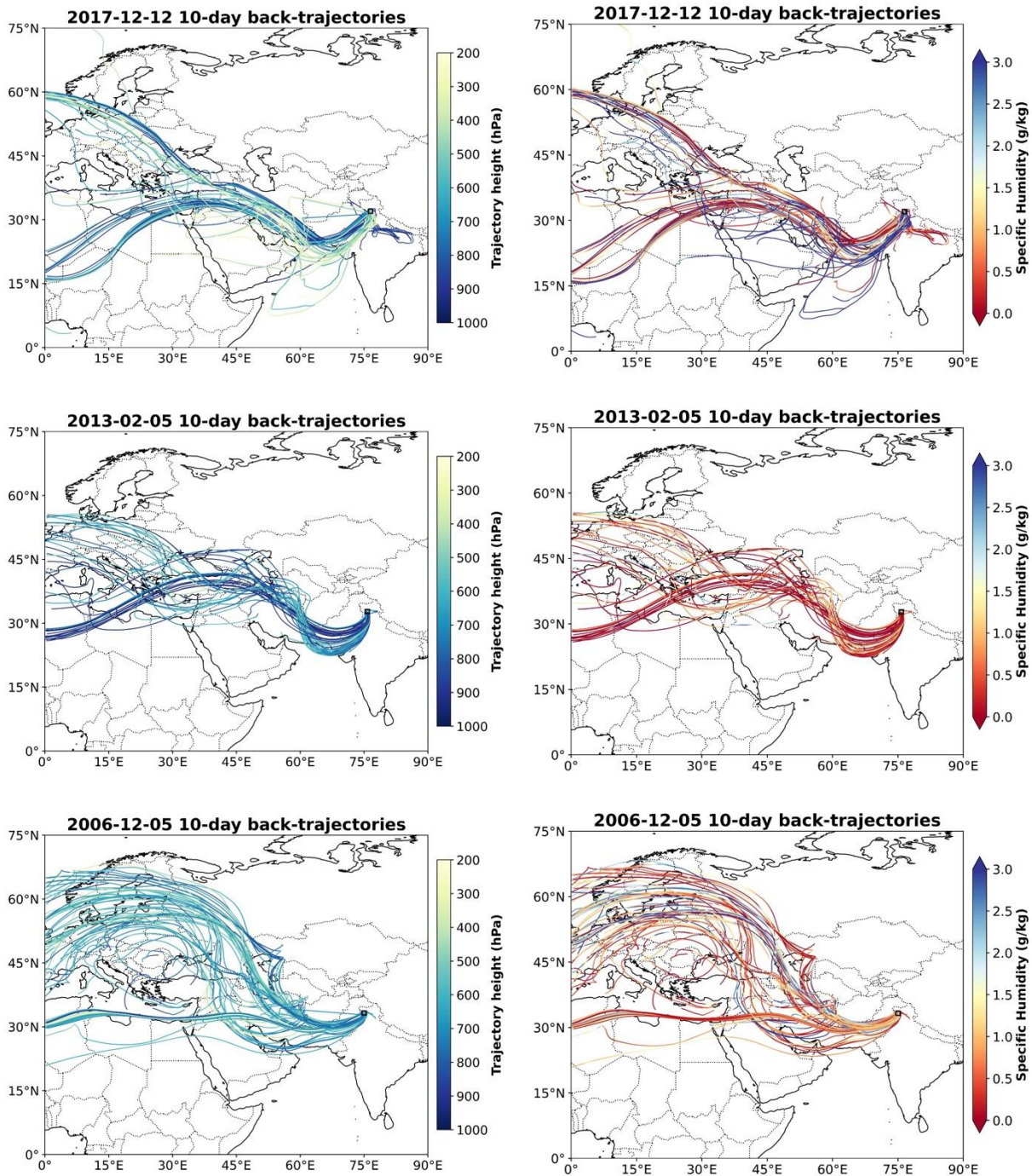
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**Figure 6:** Wintertime composite anomalies for Eddy Kinetic Energy (a-b; unit:  $m^2s^{-2}$ ) and, Baroclinic instability index (c-d) in ERA5 and IMDAA datasets during 1979- 2019.



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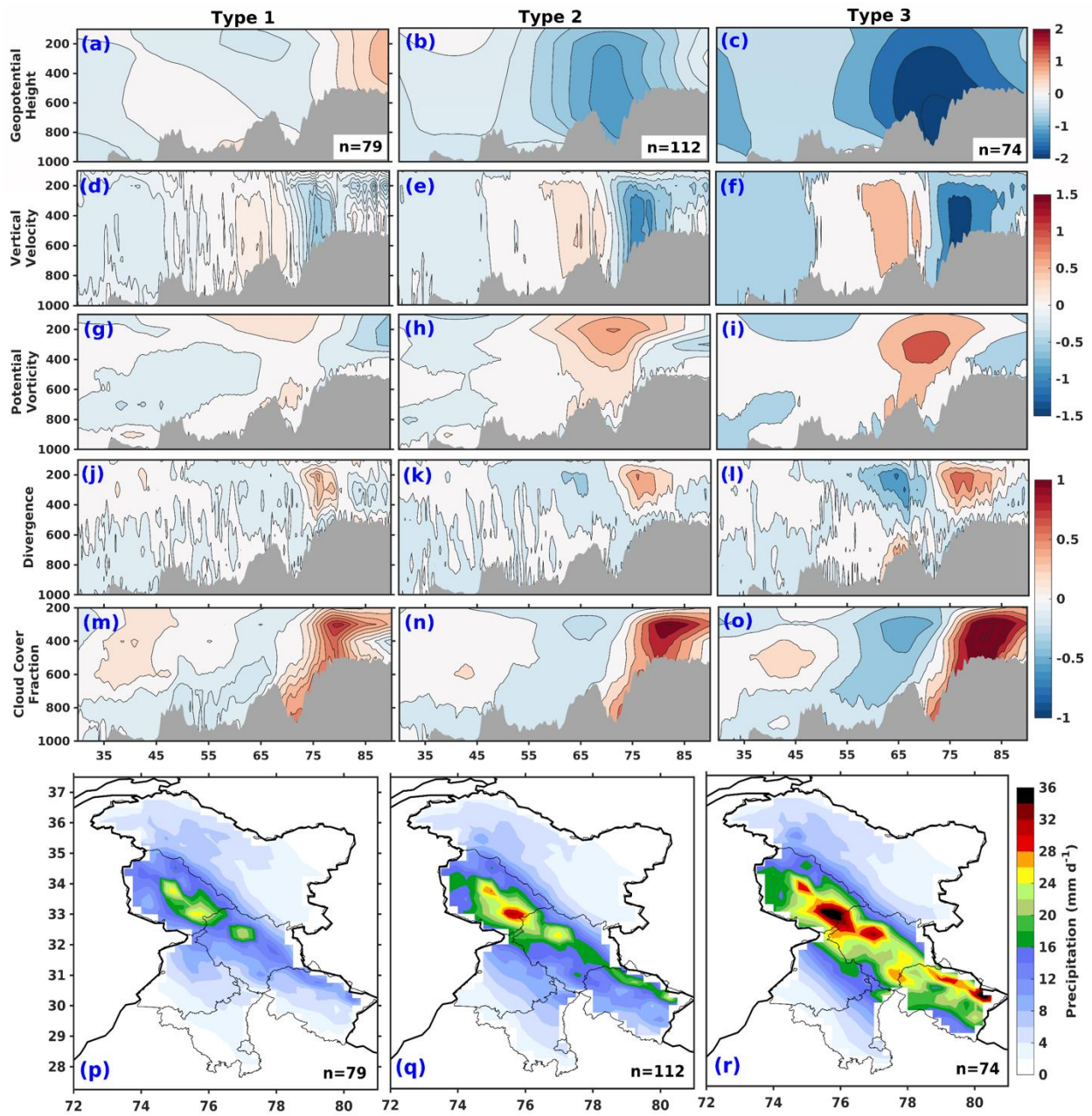
**Figure 7:** Composite anomalies (extremes minus non-extremes) for vertically integrated moisture transport (VIMT; shaded (magnitude) and vectors, unit:  $\text{kg m}^{-1} \text{s}^{-1}$ ) between surface and 300 hPa and, correlation between VIMT and frequency of extremes exceeding 95<sup>th</sup> percentile over the box (72.5°E-80.5°E, 27.5°N-37.5°N) in ERA5 and IMDAA for 1979-2019 during the winter season.



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1118 **Figure 8:** Forty parcels back trajectories (first column) initialized in atmospheric columns for selected  
 1119 winter EPEs, equally spaced between 925 and 500 hPa pressure levels. The trajectories are computed  
 1120 backward for 10 days according to the local and contemporaneous ERA-5 winds. Specific humidity  
 1121 corresponding to respective case studies have been provided in the right panel. The columns are located  
 1122 at (a) 32.25°N, 76.5°E (first row); (b) 32.75°N, 75.75°E (second row) and (c) 33.25°N, 74.75°E, (third row)  
 1123 for the extreme precipitation events on 12 December 2017, 05 February 2013 and 05 Dec 2006,  
 1124 respectively.

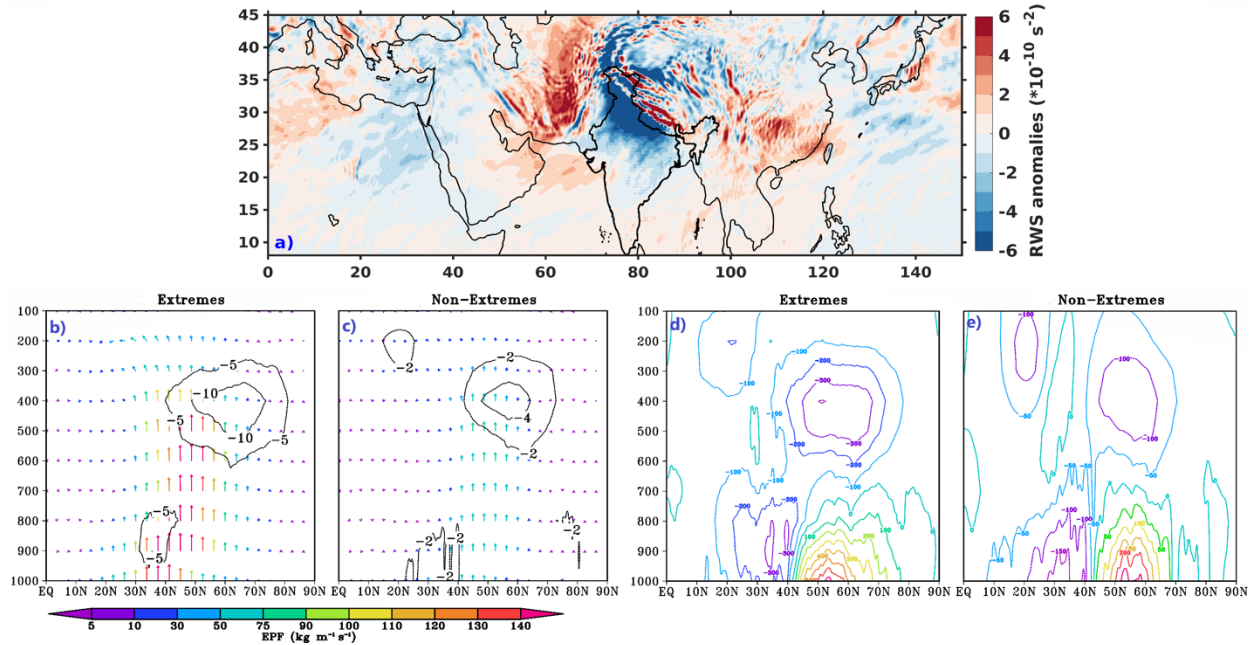




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 1126 **Figure 9:** Classification of normalized anomalies of different dynamic and thermodynamic variables for  
 1127 precipitation extreme days (exceeding 95<sup>th</sup> percentile threshold) based on a three-partition k-means  
 1128 clustering algorithm (a-o) and corresponding classification of precipitation intensity during extremes (p-r)  
 1129 based on ERA5. Populations belonging to each cluster are given in respective panels.

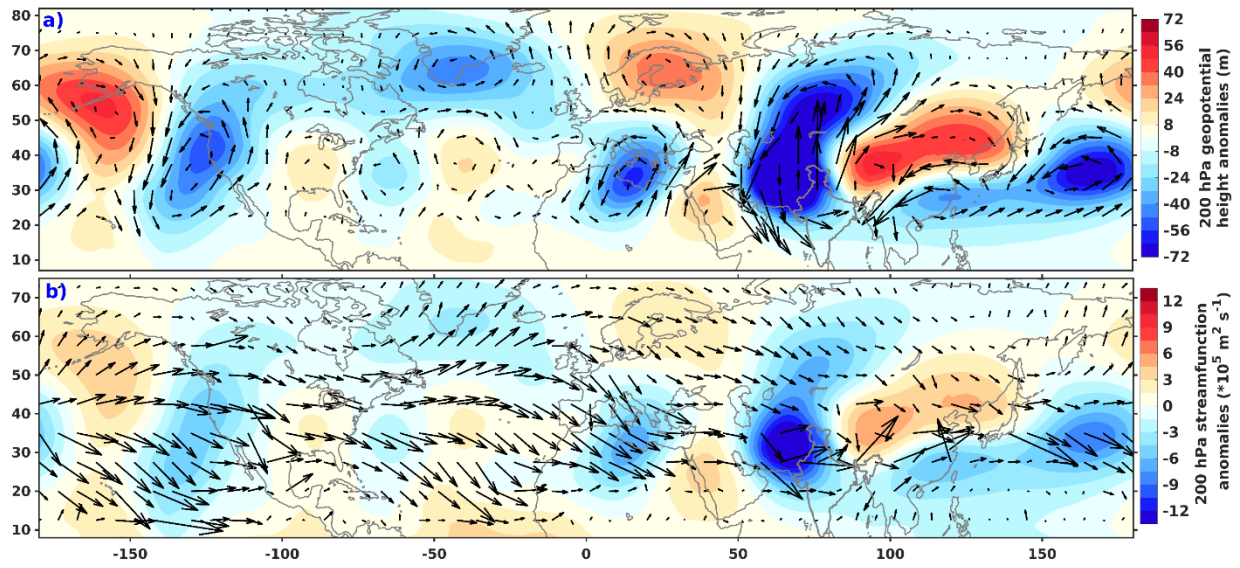
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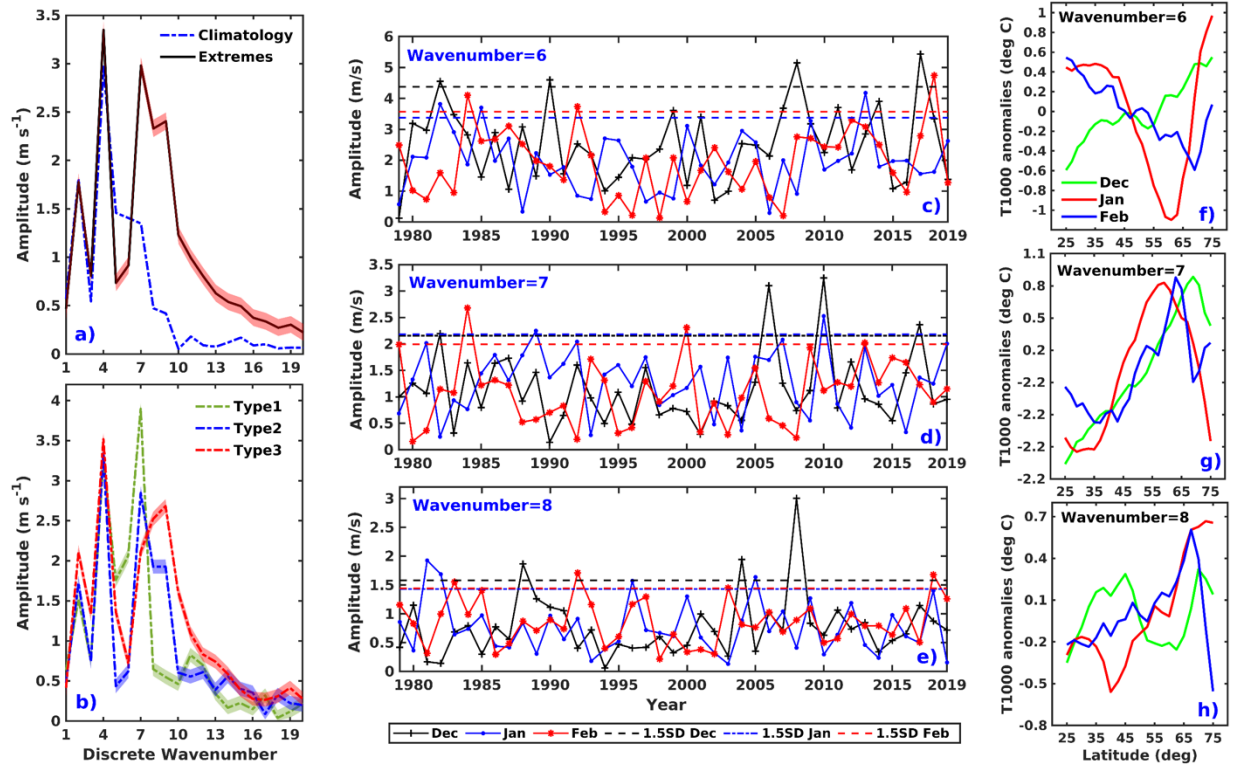
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**Figure 10:** Winter composites (extremes minus non-extremes) for Rossby wave source anomalies (a; unit:  $s^{-2}$ ) and, latitude-pressure cross-sections for zonally averaged Eliassen-Palm (EP) flux (b-c; vectors; unit:  $m^3s^{-1}$ ), westerly wind acceleration (b-c; contours; unit:  $ms^{-2}$ ) and, EP divergence (d-e; unit:  $ms^{-2}$ ) in ERA5 during 1979-2019.



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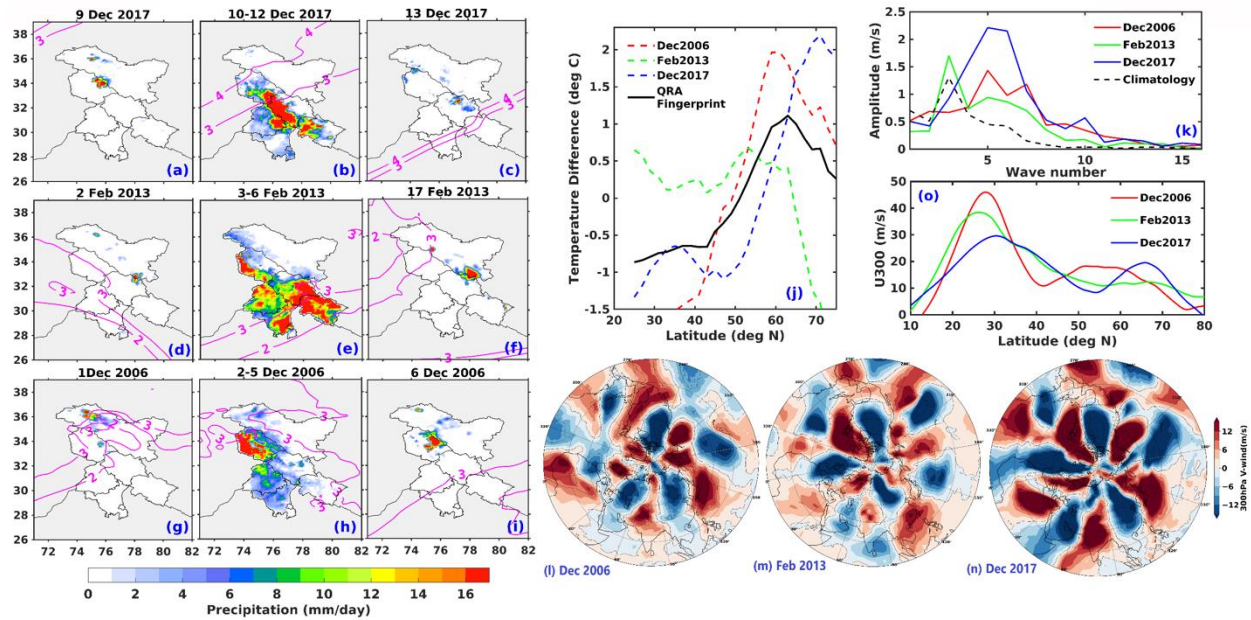
**Figure 11:** Spatial distribution of composited anomalies of upper-tropospheric (200 hPa) geopotential height (a; shaded; unit: m), wind vectors (a; unit:  $\text{m s}^{-1}$ ), stream function (b; shaded; unit:  $\text{m}^2 \text{s}^{-1}$ ) and wave activity flux (b; unit:  $\text{m}^2 \text{s}^{-2}$ ) in ERA5 dataset during 1979-2019.



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**Figure 12:** Zonal wave number spectra for the 300 hPa meridional wind averaged zonally over 30°-45°N for all identified precipitation extreme day composites in comparison to climatology (a) and; for extreme days composites associated with different clusters (b) in ERA5 dataset between 1979-2019. The filled area around the lines represents a 99% confidence interval. Interannual trends for the amplitudes of monthly mean Fourier components of the zonal wave spectra during DJF months associated with wave numbers 6/7/8. The dotted lines indicate 1.5SD amplitudes for different months (c-e). QRA fingerprint evaluated using 1000-hPa meridional temperature anomalies during DJF months associated with wave numbers 6/7/8 for the years identified from 1.5SD amplitudes in Fourier time series (f-h) and, zonally averaged 300 hPa zonal wind profiles for the same (i).





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**Figure 13:** Spatial distribution of precipitation anomalies (shaded) before, during, and after the occurrence of precipitation extremes for individual case studies (10-12 December 2017(a-c), 3-6 February 2013 (d-f) and, 2-5 December 2006 (g-i)) from GPM-IMERG and, 300 hPa isentropic potential vorticity (a-i, contours; unit: PVU) for the events (before, during, and after) from ERA5. 1 PVU (potential vorticity units) here is equivalent to  $10^{-6}$  km<sup>2</sup>/kg/s. QRA fingerprint (j) evaluated using 1000 hPa meridional temperature anomalies for individual case studies and their composite (black line). Zonal wave number spectra for the 300 hPa meridional wind averaged zonally over 30°-45°N for all three events in comparison to climatology (k), composited 300 hPa meridional wind anomalies for the selected events (l-n) and, zonally averaged 300 hPa zonal wind profiles for the same (o).