

## Eddy transport, wave-mean flow interaction, and Eddy forcing during the 2013 Uttarakhand extreme event in the reanalysis and S2S retrospective forecast data

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### Eddy transport, Wave-mean flow interaction, and Eddy forcing during the 2013 Uttarakhand Extreme Event in the Reanalysis and S2S Retrospective Forecast Data

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1	Eddy transport, Wave-mean flow interaction, and Eddy forcing
2	during the 2013 Uttarakhand Extreme Event in the Reanalysis
3	and S2S Retrospective Forecast Data
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21

### **Abstract**

22 In this study, to explore the wave-mean interaction during the monsoon season, we investigate (a) the potential role of transient eddy forcing and the wave-mean interaction on 23 the monsoon weather during the June 2013 heavy rainfall event over the Himalayan regions 24 (especially the Uttarakhand State of India and the nearby regions) and, (b) how they are 25 captured in a set of operational models. Some studies have pointed out how prolonged breaks 26 can occur due to extratropical trough incursions. However, there is a lack of clarity on how 27 transient eddy forcing associated with such interactions can lead to modulation of monsoonal 28 circulation or whether such interaction can lead to heavy rainfall events. 29 E-vector fields are analyzed to quantify the eddy forcing from extratropical transient 30 eddies and the feedback mechanism between transient eddies and the mean flow during June 31 32 2013. Analysis reveals that along with local factors (orography, moisture convergence), the large-scale heavy rainfall event over the Uttarakhand region during 16-17 June 2013 was 33 influenced by eddy forcing due to the intrusion of extratropical Rossby waves over the Indian 34 region. The location of eddy affects the location of regional occurrence of the eddy-mean 35 interaction. Model hindcast analysis results suggest that operational models cannot forecast 36 37 the upper-level eddy forced circulation patterns, and the improper representation of the Evector divergence field leads to the underestimation of intensity and the spatial pattern of 38 rainfall. 39

41

### 42 1. Introduction

It is well known that extratropical-tropical (E2T) interaction can cause prolonged breaks 43 (low rainfall spells) over the Indian region due to intrusion of extratropical troughs and ridges 44 during the monsoon season (Ramaswamy, 1962; Raman and Rao, 1981; Krishnan et al., 45 2009; Fadnavis and Chattopadhyay, 2017). However, in recent years, studies have shown that 46 high-intensity rainfall events over the Himalayan region are sometimes associated with the 47 coexistence of low-frequency monsoon intraseasonal oscillations and extratropical eddies (or 48 Rossby waves or Western Disturbances). The eddy transport and fluctuations are evident 49 through the associated heat and momentum transport towards or out of the tropics during the 50 summer monsoon, as shown in Kalshetti et al. (2021). Disturbances originating in the 51 extratropics can locally shift the jet stream equatorward and thus disturb the synoptic setup of 52 the Indian summer monsoon. These intrusions are supposedly linked to high-intensity rainfall 53 events like the 2010 Pakistan and 2013 Uttarakhand floods (Hong et al., 2011; Lau and Kim, 54 2011; Joseph et al., 2015; Vellore et al., 2016a; Sooraj et al., 2020) and many similar weather 55 events. 56

57 How does the extratropical wave intrusion create high-intensity rainfall over Northern 58 India, especially over the Himalayan region? The high-intensity rainfall event can be 59 dynamically defined as zones of strong vertical velocities with ample moisture supply from

the boundary layer or lower troposphere (such as during monsoon time), which is likely due 60 to strong ageostrophic components (Bohlinger et al., 2019). Can a Rossby wave intrusion 61 cause this high-intensity rainfall or other types of extreme events, e.g., heatwaves? Classic 62 numerical experiments such as from Hoskins and Karoly (1981) and some recent case 63 studies based on Eurasian blocking suggests that the 2010 western Russian heatwave was 64 forced by downstream Rossby wave propagation (Trenberth and Fasullo, 2012), with high-65 frequency Rossby waves also propagated southward during the end of July (Lau and Kim, 66 2011). Thus, Rossby waves are often a precursor of extreme heatwave events. Similarly, the 67 leading trough of the Rossby wave train triggered upward motion to the east, favoring deep 68 convection leading to extreme rainfall event-related flooding over Pakistan during 2010 69 (Hong et al., 2011; Lau and Kim, 2011). Likewise, extratropical Rossby wave-linked 70 convection is also reported in the tropical Pacific (Kiladis, 1998). Heavy rainfall, especially 71 extreme events, is associated with mesoscale convective systems. Several studies on 72 midlatitude convection showed that midlatitude Rossby waves are reinforced by mesoscale 73 systems (Lillo and Parsons, 2017; Parsons et al., 2019), which create downstream forced 74 75 response. Some studies have also shown how mesoscale convective systems over the Pacific warm pool region get reinforced by Rossby-kelvin waves(Houze et al., 2000), indicating the 76 role of large-scale waves in local mesoscale precipitating systems. In addition to waves, 77 studies have also identified thermal forcing and background moist processes over the arid 78

79	region of Indo-Pakistan in the 16 to18 days prior and have shown them to be precursors for
80	extreme rainfall events over the monsoon core zone (Sooraj et al., 2020).

As mentioned in the last paragraph, the state of Uttarakhand in India witnessed heavy 81 rainfall during 15-17 June 2013 with widespread reporting of a "cloudburst"-like scenario 82 over the Kedarnath region in Uttarakhand. Some studies have demonstrated the role of 83 mesoscale convective systems in cloudburst situations(Parida et al., 2017). Other studies 84 suggest high-intensity rainfall events such as those that occurred over the Uttarakhand region 85 during June 2013 result from an occluded frontal system that developed in response to a 86 western disturbance and monsoon low (Chevuturi and Dimri, 2016). The validity of the 87 frontal theory of ageostrophic uplift requires the validity (i.e., finiteness) of the ratio NH/f (N 88 is Brunt-Väisälä frequency, f is the Coriolis parameter, and H is the depth of the fluid) as 89 described in Hoskins (1982). In the tropics, the validity of this ratio is not well defined due to 90 the smallness of f. Also, as (Charney 1969) suggested, such large-scale lifting in the stably 91 stratified tropical climate is not easily (or generally) possible due to lack of vertical coupling 92 unless it occurs in the regions of active tropical deep cumulus convection. 93

In addition to such localized extreme events during 16-17 June, heavy to very heavy rainfall occurred during this time over larger regions (in the synoptic scale, e.g., refer fig.2 of Joseph et al., (2015) or Fig.12). The coexistence of synoptic-scale heavy rainfall and Rossby wave intrusion may indicate that the transient eddies may be a dominant factor in comparison 98 with other factors like local destabilization due to latent heating, which is already reported

99 (e.g., Fig. 9b of (Wills and Schneider, 2018).

This eddy (i.e., transient synoptic disturbance) perspective has also been explored in 100 several studies. Vellore et al. (2016b) and Hunt et al. (2018a, 2018b), using cluster composite 101 analysis, looked into several extratropical intrusions of western disturbances in a detailed 102 manner and described the presence of monsoon low-pressure systems and western 103 disturbance as a necessary condition for heavy events. Another recent study (Hunt et al., 104 2021) explained the extreme events in terms of the coexistence of monsoon low-pressure 105 systems and western disturbances. They found that high-intensity precipitation occurs as a 106 result of their interactions, but their analysis uses the perspective of Hanley et al. (2001), 107 which assumes the pre-existence of an intense cyclonic storm (with substantial azimuthal 108 eddy flux transfer, EFCs which is a measure of local eddy forcing) co-located with the jet 109 stream and that this storm interacts with the upper-level troughs when the eddy-induced 110 azimuthal flux transfer measured by EFCs cross certain thresholds and provide necessary 111 eddy forcing. This EFC-based criterion is, however, not quantitatively verified in these 112 studies. 113

The studies described above indicate that the orography, monsoonal moisture flow, stability, and waves/eddies can contribute to the extreme as well as heavy rainfall event, but do not go into explaining this monsoon midlatitude eddy interaction scenario as an eddy

117	forcing problem on the mean flow (i.e., eddy-mean interaction perspective), and hence,
118	though the earlier analysis highlights exchanges, they neglect a transient wave forcing
119	perspective (e.g., refer eq. 1 in sec.3). In the current analysis, we will focus on how (or
120	whether) wave-mean interaction can be used to explain ageostrophic vertical velocities and
121	rainfall that occurs during such events. Such interactions perspective are often useful to study
122	the mid-latitude low-frequency oscillations (Jin, 2010).

Models fail to predict the rainfall pattern and amplitude over Uttarakhand with sufficient 123 lead-time Fig.1 shows that several operational dynamical models lack skill in predicting the 124 rainfall event over Uttarakhand even at a one-week lead-time. Several models have these 125 problems with events related to Rossby waves (e.g., forecast bursts as referred in (Lillo and 126 Parsons, 2017). In addition to understanding the eddy-mean interaction during the intrusion, 127 the eddy forcing approach can be useful if we want to know why the dynamical models fail to 128 forecast such events at sufficient lead times for issuing alerts, especially in the weather to the 129 extended-range forecast time scale. We hypothesize that the eddy-mean interaction 130 perspective can be helpful in understanding the skill (or lack of it) in dynamical models. 131

The Uttarakhand extreme event of 2013 and the high-intensity rainfall over large regions has been studied by many researchers and is a well-known event in which tropicalextratropical interaction is well documented (Dube *et al.*, 2014; Vellore and Jayant, 2014; Joseph *et al.*, 2015; Pattanaik *et al.*, 2015; Singh and Chand, 2015; Vellore *et al.*, 2016a; Kaur and Gupta, 2017). Hence, owing to the good documentation of linear perspective, the

current paper selects this event and discusses the eddy-mean interaction perspective. We 137 elaborate on two aspects: (a) whether wave-mean interaction was significant during this event 138 in addition to the earlier linear theories used to explore such events and (b) based on this 139 transient eddy wave mean interaction formulation, whether such steps can be retraced or 140 evaluated in the dynamical models to understand the lack of skill in these models when 141 forecasting such events. The study is arranged as follows: Sec.2 describes the study area, data 142 used, and formal methodology. In sec.3, we computed the transient eddy forcing during the 143 extreme event using the *E-vector* approach (Andrews and McIntyre, 1976; Hoskins et al., 144 1983), which explains the eddy dynamical pathway for such events near subtropical jet 145 streams. Sec.4 describes the application of the E-vector approach while describing the life 146 cycle of tropical-extratropical interaction while describing the intrusion event during the 147 Uttarakhand heavy rainfall event (June 2013). Sec.5 describes the operational forecasts of the 148 event, and the results are discussed, summarized, and concluded in Sec.6. 149

- 150 2. Study area, Data, and Methods
- 151 *2.1. Study area*

The natural disaster, in the form of landslides and flash flooding, occurred due to heavy precipitation over the Uttarakhand region (29°-31°N, 78°-81°E), India on 16-17 June 2013. Uttarakhand is part of the region of complex orography along the Western Himalayas. The Western Himalaya region surrounds snow-covered peaks, crest, glaciers, valleys, and perennial river basins (Parida *et al.*, 2017).

### 157 *2.2. Observation Data*

The atmospheric (retrospective) reanalysis dataset from the fifth-generation European 158 Centre for Medium-Range Weather Forecasts (ECMWF) Atmospheric Reanalysis of the 159 global climate (ERA5) is used as a proxy for observation of dynamical variables. ERA5 data 160 are produced by the Copernicus Climate Change Service (C3S) at ECMWF. ERA5 climate 161 data is available in a  $0.25^{\circ} \times 0.25^{\circ}$  grid from 1979 to within five days of real-time (Hersbach 162 et al., 2020). The primary data used in this study are zonal (u, ms<sup>-1</sup>), meridional (v, ms<sup>-1</sup>), 163 vertical ( $\omega$  Pa s<sup>-1</sup>) wind, temperature (K) at 200hPa, and geopotential height (m) at 500hPa. 164 The observed precipitation is obtained from Tropical Rainfall Measuring Mission (TRMM) 165 Multi-satellite Precipitation Analysis (TMPA, 3B42) (Huffman et al., 2007). 3B42 data 166 contains a gridded, satellite-based merged infrared precipitation data (mm/hr), with a 3-hour 167 temporal resolution and a 0.25-degree spatial resolution. We have converted the sub-daily 168 data to daily data. 169

170 *2.3. Models* 

The Indian Institute of Tropical Meteorology (IITM) extended range forecast (IITM-ERPAS) runs (Abhilash *et al.*, 2014; Sahai *et al.*, 2019), UKMO (United Kingdom Met Office), and ECMWF (European Centre for Medium-range Weather Forecasts) forecast runs are the S2S reforecast model runs used in this study. All the observed and model data set regridded to 1° ×1° spatial resolution and daily temporal resolution with appropriate pre-

176	processing. Except for IITM-ERPAS (Abhilash et al., 2014), the UKMO and ECMWF
177	forecast data were accessed through the S2S "instantaneous and accumulated" database
178	maintained by ECMWF (Vitart et al., 2017). ERPAS data is obtained from the IITM-IMD
179	operational database maintained at IITM. Each S2S model has a control member (using
180	unperturbed initial conditions) and several perturbed members produced for sampling
181	uncertainty in the initial conditions. In the present study, the model forecast is taken from the
182	closest available initial condition on 9th June 2013 for ERPAS and UKMO and for 10th June
183	2013 for ECMWF when the forecast of 16-17 June lies in the synoptic range. Spatial plots are
184	shown for the ensemble mean (IITM-ERPAS-16-member mean; UKMO-7 member
185	(including one control member); ECMWF-11 member (including one control member) unless
186	stated otherwise.
187	3.0 A schematic description of the life cycle of extratropical intrusion
188	The extratropical eddies are associated with colder extratropical air masses and stronger
189	westerly momentum in the upper levels than tropical monsoonal flow. The transport of this

westerly momentum in the upper levels than tropical monsoonal flow. The transport of this eddy momentum and heat flux over the Indian region in the presence of low-frequency monsoonal background (monsoon intraseasonal oscillations and low-pressure systems (Goswami, 2012)) can significantly impact the local flow as it imparts additional heat and momentum in the local budget terms (cf. eq. one and eq.2 in sec.3.2). The event life cycle can thus be hypothesized to have three major stages: (i) southward *digging* (or intrusion with amplification) of troughs associated with a meandering of the subtropical jet stream, (ii) transient eddy forcing onto the mean flow, and (iii) reversal of jet stream (or weakening of
southward intrusion) position with the restoration of monsoon flow. These are described
below.

199

200

# 3.1 Linear Process: Intensification of digging trough over the Indian Region during the monsoon season

201 Rossby waves show meridional propagation, but the southward intrusion of the Rossby wave in the tropical region is generally restricted due to the presence of critical latitudes where the 202 phase velocity of the wave approaches the zonal (westerly) wind speed, and the waves are 203 204 absorbed along the critical latitudes as it propagates equatorward (Karoly and Hoskins, 1982). Normally Rossby waves break, and mixing of potential vorticity (PV) occurs with the decay 205 of the wave leading to mid-latitude to tropical exchange (Homeyer and Bowman, 2013). 206 However, Rossby waves can dig deep south into the tropical region due to the ducting effect 207 or southward extension of westerlies over a certain region (Webster and Holton, 1982). The 208 southward shifting/meandering of the westerly jet stream due to the presence of the blocking 209 high can provide such ducts where Rossby waves can propagate into the tropics. At the same 210 time, cyclonic Rossby wave breaking can be prevented, and it can intensify over the Indian 211 region in the presence of monsoonal instabilities. It can be seen as follows: 212

213 Consider the conservation of PV in the absence of diabatic heating:  $\frac{\zeta + f}{\frac{\partial p}{\partial \theta}} \approx Const$ 

214 If a cyclonic phase Rossby wave (positive vorticity) comes into a region of higher thickness (

215  $\frac{\partial p}{\partial \theta}$ ) and f decreases, relative vorticity has to increase. For more detailed explanations of the

216	PV aspects of synoptic developments, the reader refers to case studies discussed in Hoskins
217	(1997). Thus, the cyclonic vorticity induced by the Rossby wave can intensify (or at least its
218	amplitude does not decay) depending on the strength of the denominator, as a southward
219	extension means a decrease in $f$ . Typically such thickness would increase in the presence of
220	strong heating. Hence the Rossby wave would strengthen and not break or weaken in the
221	presence of background instability that modifies. Such increased background thickness can be
222	provided by latent heat release associated with modes of subseasonal variability when the
223	organized convection stays over the foothills of the Himalayas.

The intruded (or possibly intensified) Rossby wave can create a digging trough, leading to indirect ageostrophic circulation with the high vertical velocity at the jet exit region (e.g., Chapter 2.3 and Fig. 2.7 of Lackmann (2012). In addition to such direct effects, the strengthening or weakening of the westerly or the Rossby wave depends on the growth of eddy instability, which can extract energy from the mean flow (Held and Phillips, 1989). The response of the mean state due to transient eddy forcing is essential to control the growth or decay of the Rossby wave. This nonlinear process is discussed next.

### 231

232

3.2 Nonlinear process: the transient eddy forcing of the mean flow and Monsoon time mean background

There are numerous studies demonstrating empirical evidence of the role of the 233 transient eddies in maintaining the time-mean flow and how eddies extract energy from the 234 mean flow or give energy to the mean flow (Lau and Holopainen, 1984; G. Branstator, 1995; 235 Jin, 2010; Tan *et al.*, 2014). In particular, a recent study (Wills and Schneider, 2018) showed 236 transient eddy feedback for Rossby waves over the orographic region is more important than 237 latent heating. The eddy mean interaction for the horizontal momentum equations, assuming 238 zonal wind (u) and meridional wind (v) compartmentalized into mean and transient eddy 239 components:  $(\mathcal{V} = \overline{\mathcal{V}} + \mathcal{V}', \mathcal{V} = (u, v))$  (Hoskin et al. 1983; James 1994; Williams et al. 2007) 240 is given by: 241

242 
$$\frac{\overline{D}\overline{u}}{Dt} = f + \nabla . E - \dots - (1a)$$

243 
$$\frac{\overline{D}\overline{v}}{Dt} = -fu_{am} - (\overline{u'v'})_x$$
-----(1b)

 $\frac{\overline{D}}{Dt} = \overline{u}\frac{\partial}{\partial x} + \overline{v}\frac{\partial}{\partial y}$ , ()<sub>x</sub> =  $\frac{\partial}{\partial x}$ ,  $\nabla$  is the horizontal gradient operator and f is the Coriolis 244

parameter.  $V_{am} = (u_{am}, v_{am})$  is the "modified ageostrophic" wind given by  $\overline{v_{am}} = \overline{v} - \overline{v}$ 245

 $f^{-1}k \times \nabla(\overline{\phi} + \overline{\nu'}^2)$ . Refer to Appendix A of Hoskins et al. (1983) for more details. 246

*E* in the equation. (1a) is the so-called E-vector defined as: 247

248 
$$E = \left(\overline{v'^2 - u'^2}, \overline{-u'v'}\right).....(2)$$

Where  $\overline{v'^2 - u'^2} = Ex$ , Eddy anisotropy or asymmetry,  $\overline{-u'v'} = Ey$ , eddy 249 momentum flux transport (EMF)

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The E-vector is a generalized form of Eliassen-Palm flux (Andrews and McIntyre, 1976) and can be used to estimate the local impact of transient eddies (Hoskins *et al.*, 1983; Trenberth, 1986; James, 1994).

The above relations suggest that the E-vector flux divergence can change zonal and 253 meridional wind (assuming other factors like friction are negligible in the upper level), i.e., 254 eddies can force the mean flow. Divergence of the E-vector can cause acceleration of mean 255 flow, and convergence can decelerate the mean flow. In a monsoonal setting, if we take the 256 terms in the LHS of Eq.1 as representing the mean monsoonal quasi-stationary background, 257 the second term in the RHS can be interpreted as the eddy forcing term arising due to 258 extratropical intrusions or similar low-pressure systems. Depending on the sign of the eddy 259 convergence, the eddy can either grow (or extract energy from mean flow) or they can 260 dissipate by supplying energy to the monsoonal mean flow (also refer to (Held and Phillips, 261 1989). To understand and estimate the impact of this eddy forcing during the Uttarakhand 262 event, we will use this E-vector formalism. 263

264 The E-vector is also related to transient eddy vorticity flux convergence and divergence. It

265 can be shown (Hoskins *et al.*, 1983) that: 
$$\nabla . \overline{u'\zeta'} \approx \frac{d}{dy} (\nabla . E)$$
.....(3)

Then, from Eq.1 and Eq.3, we can see that an E-vector framework provides a firsthand idea of the transient eddy forcing and zonal asymmetries imparted by the *transient* eddies (Hoskins *et al.*, 1983; Trenberth, 1986) in the momentum and the vorticity equations.

Also, plotting of E-vectors, i.e. (Ex, Ey), can give an idea about the direction and location of eddy convergence. The E-vector approach can be used to study both zonally asymmetric tropical (Leroux *et al.*, 2010) and extratropical transport patterns (Novak *et al.*, 2015). The above discussion is only based on the mechanical effects of eddies. A complete quasigeostrophic discussion including the thermal effects of eddies by defining a third component in E-vector using temperature (T) and meridional wind ( $\sim \overline{v'T'}$ ) can also be made analogously but is beyond the scope of our study.

276 **3.3** *Extratropical intrusion and heavy rainfall events* 

The next question is, how are the extratropical intrusions or the eddy forcing are 277 linked to the high-intensity rainfall events? We can have two possibilities based on the above 278 two sections (3.1 and 3.2). First, a convergence of eddy flux can be interpreted as a positive, 279 i.e., cyclonic curl directly introduced in the mean flow (eq.12 and 13 of Williams et al., 280 (2007). In a monsoon environment, this can invigorate the existing monsoonal convergence 281 directly over a region where the curl is created (such as when a monsoon trough exists over 282 the foothills of the Himalayas). The second possibility is through an ageostrophic secondary 283 circulation mechanism. The generation of cyclonic (anticyclonic) vorticity in the lower 284 troposphere (transverse circulation dynamics) is associated with the formation of surface 285 lows and indirect circulation (e.g., Fig.2 of Hoskins 1982). Such circulation will lead to the 286 generation of ageostrophic vertical velocity, and depending on local factors (like orography 287

and moisture), dramatic intensification can occur over the elevated topographies (Uccellini 288 and Johnson, 1979; Vellore et al., 2016a; Hunt et al., 2021), leading to extreme precipitation, 289 as a result of a jet streak excitation as shown by Hunt *et al.*, (2021). This ageostrophic motion 290 associated with the indirect circulation can extract kinetic energy from the mean flow in the 291 downstream region of the jet exit, where it is super-geostrophic (Lau, 1979). Extreme 292 precipitation can result from strong ageostrophic vertical upward motion restoring static 293 stability. Based on Orlansky and Sheldon (1995), the quasigeostrophic omega equation can 294 be written as: 295

296 
$$f_0^2 \frac{\partial}{\partial z} \left( \frac{1}{\rho \partial z} \rho w \right) + N^2 \nabla^2 w = s_g + s' - \dots - (4)$$

Where *w* is the vertical velocity,  $\rho$  is the density,  $f_0$  is the Coriolis parameter,  $N^2$  is the Brunt-Vaisalla frequency and *s'* is the source term involving friction, diabatic heating, and meridional gradient of Coriolis parameter.  $s_g$  is the source term involving geostrophic quantities (denoted by *g* indicated in the suffix), which Sutcliffe approximated as:

301 
$$s_g = 2|f \partial \boldsymbol{\mathcal{V}}_g / \partial z| \frac{\partial \zeta_g}{\partial l}$$
,-----(5)

where l is a unit coordinate aligned along the thermal wind direction and  $v_g$  is the geostrophic wind. From Eq (4), (5), (3), and (1), it is clear that  $\zeta_g$  as well as  $\partial \mathcal{V}_g/\partial z$  can be forced/modulated by eddy forcing through the generation of vertical velocities.  $\nabla \boldsymbol{E}$  forces the zonal mean wind, which then changes the vertical shear, subsequently changing the source term  $s_g$ . A

large increase in vertical shear due to E-vector divergence would then lead to a proportional 306 increase in vertical velocity. Under the right conditions (e.g., moisture supply), increased 307 vertical velocity can result in an increase in rainfall. The jet will move back to the original 308 position after the Rossby wave has propagated out (or dissipated). The cloud-free conditions 309 restore the horizontal meridional (north-south) gradients of temperature, forcing the jet to 310 move back to its normal position. The mechanisms explained in this section are summarized 311 in **Fig.2**. The theoretical mechanism proposed here can be verified by diagnosing the direct 312 (mean flow modification, cf eq. 1) and indirect (transverse circulation) mechanism based on 313 the E-vector field will be explored in the next session. 314

#### 4.0 The lifecycle of extratropical intrusion during the 2013 Uttarakhand extreme event 315

In the context of the sequence of events described in sec.3, the following section 316 describes the precipitation and circulation plots to quantify the linear and nonlinear 317 perspectives during this event. To understand the nature of the extratropical intrusion, we plot 318 the first two components of E-Vectors (eddy asymmetry term  $(\overline{u'^2} - \overline{v'^2})$  and the eddy 319 momentum flux transfer term (u'v') or the EMF term) area averaged over a domain north of 320 the Uttarakhand (30°-40°N;75°-82°E) region in **Fig.3a**. The primes are defined with respect 321 to the monthly mean for June 2013. The plot shows that there is strong eddy asymmetry  $\overline{v'^2}$ 322  $\gg \overline{u'^2}$  during 16-17 June 2013 and strong northward transfer of EMF. Such meridional 323 elongation is significant and consistent with theory in the sense that it indicates southward 324

phase propagation associated with northward momentum transfer (e.g., sec. 3b of (Waterman
and Hoskins, 2013) A similar EMF flux but weak eddy asymmetry amplitude are seen during
7-9 June. This implies that both flux transfer and eddy asymmetry could play an essential role
during this event. This meridional intrusion will be further discussed in detail in sec.4.2.
Next, we will see the synoptic situations during this time.

330

331

## 4.1 Precipitation and circulation during the event: synoptic-scale nature of the event

Fig.3a shows the strengthening of the eddy asymmetry and EMF index (first and the 332 second components of the E-vector), which helps to understand the low-frequency intrusion 333 over the Indian region (Kalshetti et al., 2021). To understand the synoptic situation, we first 334 analyze the precipitation during June 2013. Fig. 3b shows the northward propagation based 335 on 20-80 days Lanczos-filtered TRMM precipitation data during June 2013. It shows clear 336 intraseasonal northward propagation in this band. Around 13-17 June, the rainfall band 337 strengthened due to the expected monsoon progression after the onset phase (also refer to 338 Fig.1). Similarly, Fig.4 shows the evolution of the rainfall and the 200hPa wind pattern. The 339 meandering contours confirm the deeper southward movement of the jet stream staying 340 southward of its climatological mean position during summer (~35N). The temporal 341 evolution plots in Fig.4 show how the blob of precipitation evolves and decays over the 342 Uttarakhand region. It is clear from this plot that the rainfall band spans over larger areas 343

344	spanning a few thousand square kilometers, indicating that the event which caused the high-
345	intensity rainfall is not a purely local extreme event phenomenon and is likely to be
346	correlated with large-scale structures. This large-scale nature is further confirmed from the
347	vorticity plots in Fig.5. The temporal evolution on 16th June and 17th June shows cyclonic
348	vorticity advection with strong vorticity bands moving eastward with the additional formation
349	of jet streaks (on 16th June, the strongest vorticity shading is in the southward descending
350	branch west of 75°E and on 17 <sup>th</sup> June it is east of 7°5E in the northward ascending branch).
351	The contours of EMF terms show strong northward (i.e., positive contours) transport of eddy
352	momentum. Thus, cyclonic vorticity advection could be responsible for creating anomalous
353	eddy momentum fluxes, and hence the Uttarakhand region shows more substantial eddy
354	forcing during this period as compared to other days in June 20013.

In addition to the extratropical system, as shown in Fig. 3, a low-pressure system was 355 also propagating over the Indian region. The low-level circulation and the sea-level pressure 356 pattern are shown in Fig.6. The strong low-level circulation is associated with strong 357 moisture inflow. Also, the surface pressure shows negative anomalies over northwest India 358 and positive anomalies over the eastern part along the Bay of Bengal. Strong northerlies and 359 anomalous high pressure and low-pressure regions are also visible over the Afghanistan 360 region, indicating the existence of upper-level waves. How is the moisture transported in the 361 upper level? We plot the 500-200hPa averaged moisture transport (*uq*, *vq*) and see the strong 362 moisture outflow from the southern side, particularly from the Bay of Bengal low-pressure 363

364	system towards the Uttarakhand region in Fig.7a-d. The north-westerly moisture transport is
365	strong in the upper level, with the dominant source of moisture being the Bay of Bengal. The
366	static stability plots (Fig.7e-h) also show a strong meridional gradient with stability
367	decreasing northwest to southeast over the location of the extreme rainfall during 16-17th
368	June 2013 (blue shading). Thus, Figs.3-7 confirm the presence of both the extratropical
369	system and the tropical system during this period with adequate monsoonal and extratropical
370	influence over the Uttarakhand region. In the next section, we will elaborate on the southward
371	extension of the extratropical intrusion
571	
372	4.2 Southward digging of extratropical troughs
372 373	4.2 Southward digging of extratropical troughs To understand the southward digging of the trough, which faces decreased stability
372 373 374	4.2 Southward digging of extratropical troughs To understand the southward digging of the trough, which faces decreased stability over the Indian region, we plot <i>Fig.8</i> . The plot has multiple variables, all averaged for the
<ul> <li>372</li> <li>373</li> <li>374</li> <li>375</li> </ul>	4.2 Southward digging of extratropical troughs To understand the southward digging of the trough, which faces decreased stability over the Indian region, we plot <i>Fig.8</i> . The plot has multiple variables, all averaged for the days 15-17 June 2013, which are described below: (a) the shading shows the vertical velocity
<ul> <li>372</li> <li>373</li> <li>374</li> <li>375</li> <li>376</li> </ul>	<ul> <li>4.2 Southward digging of extratropical troughs</li> <li>To understand the southward digging of the trough, which faces decreased stability</li> <li>over the Indian region, we plot <i>Fig.8</i>. The plot has multiple variables, all averaged for the</li> <li>days 15-17 June 2013, which are described below: (a) the shading shows the vertical velocity</li> <li>(w) with positive shading indicating upward motion, (b) the magenta curve shows the</li> </ul>
<ul> <li>372</li> <li>373</li> <li>374</li> <li>375</li> <li>376</li> <li>377</li> </ul>	<ul> <li>4.2 Southward digging of extratropical troughs</li> <li>To understand the southward digging of the trough, which faces decreased stability over the Indian region, we plot <i>Fig.8</i>. The plot has multiple variables, all averaged for the days 15-17 June 2013, which are described below: (a) the shading shows the vertical velocity (<i>w</i>) with positive shading indicating upward motion, (b) the magenta curve shows the streamlines at 200hPa, (c) blue contours shows the (positive only) relative vorticity at</li> </ul>

(e) the green curve shows the 1 PVU contour. The black contours show that the mean jet is
shifted over the Indian region and is highly meandering. This meandering indicates trough
intrusion, confirmed by the blue vorticity contour in the northern part of Pakistan and Indian
region over Kashmir and Leh-Ladakh region. The jet is pushed southward with the southward

383	extending intensified Rossby wave trough. Two anticyclones are formed on the southern side
384	of the jetstream, which is evident by the streamlines (magenta color). The deep cyclonic
385	curvature over the northern flank of India shows that the southward extension is deep with an
386	intense Rossby wave trough. The green-coloured 1PVU curve shows that the Rossby wave
387	did not break over the Indian region during this period as the breaking of the Rossby wave
388	induces higher PV (~1-2PVU or higher) over the Indian region (below 30°N(Fadnavis and
389	Chattopadhyay, 2017). The strong amplitude of anomalous relative vorticity (blue contours
390	showing positive only values) indicates that the Rossby wave is intense, with no sign of
391	weakening and breaking with jet pushed southward. At the same time, however, there is
392	upward vertical velocity (red shading)) and rainfall (Fig.4) over both the central and the north
393	Indian regions. It is interesting to note that the Rossby wave vorticity structure does not
394	directly reach the region where the heavy rainfall event occurred and does not show much tilt
395	in the zonal vertical direction until 600hPa (plot not shown). Hence, none of the above
396	discussions would indicate the rainfall over the Uttarakhand region occurs as the end product
397	of simple in-situ linear processes such as Rossby waves or monsoon low pressure-induced
398	local convergence (e.g., the linear part in Fig.2). Can this widespread rainfall be explained if
399	we assume that the ageostrophic vertical velocity is generated due to a nonlinear conversion
400	process related to downstream amplification associated with baroclinic conversion (Orlanski
401	and Sheldon, 1995) or a downstream wave-mean interaction or a combination of both over
402	the Uttarakhand region? Such nonlinear conversion processes and wave mean interaction can

403 lead to the conversion of zonal mean to eddy kinetic energy or vice versa and generate404 ageostrophic vertical velocity.

To highlight the strong vertical velocity and the ageostrophic component, the vertical 405 velocity over a region including Uttarakhand is plotted in Fig. 9 as a height-latitude plot. The 406 plot clearly shows large vertical velocity developing over the Indian region. The development 407 of vertical velocity at this scale can be contributed to by two components, one coming from 408 orographically-forced ascent and the other from wave-induced ascent associated with shears 409 (Teixeira, 2014; Cohen and Boos, 2017). As the upper-level flow is predominantly westerly 410 and largely baroclinic (i.e.,  $\partial u/\partial z > 0$ ), having developed from barotropic easterlies (i.e., 411 predominantly easterlies at all levels) before 16<sup>th</sup> June, the plot shows the development of 412 strong vertical velocity that is associated with strong vertical shear (zonal wind contours). 413 Such shear development would give rise to unstable waves and vertical velocities (contained 414 in the term sg and s' in eq.4). We also examine (not shown) the bulk Richardson index (the 415 ratio of Brunt Vaisala frequency to vertical shear). This index and also Fig.7d-f show 416 troposphere above the southern Himalayas slowly becomes unstable as the westerly shear 417 zone develops over the slopes. This generation of ageostrophic vertical velocity also indicates 418 the conversion of mean flow to eddy kinetic energy, as we demonstrate in the next section. 419

420

### 4.3 Transient Eddy Mean Interaction

421	<b>Fig.10</b> shows the E-vector divergence, i.e. $\nabla$ . ( <i>E</i> ) (shaded) and the E-vectors. The
422	mean is defined as the 21-day average from 9 <sup>th</sup> June to 30 <sup>th</sup> June, and the daily transient is
423	defined accordingly as a departure from this mean. Components of E-vectors and the
424	divergence are computed after that. The divergence plot is averaged for the period 14-18 June
425	2013. Divergence is very strong for the period 15-17 June 2013 (not shown). Also, we
426	superimposed the contour of zonal wind averaged during 15-17 June to show the location of
427	the jetstream during this period. The ∇. E-vector shadings indicate that over the Uttarakhand
428	region, $\nabla$ .E is positive, and hence the mean flow gains energy (cf. equation 1a). The vector
429	plot shows the convergence pattern of the E vector (Ex, Ey). There are two zones of
430	convergence-divergence patterns around 40°N, which are consistent with the locations of
431	blocking ridges (east and west Asian ridges). This transfer of energy is further confirmed
432	from the plot of the local transient eddy kinetic energy (EKE given by $1/2(u'^2 + v'^2)$ ) in
433	Fig.11. The black bar plot from ERA5 shows the growth of EKE during that period, which
434	reduces afterward. The growth of EKE and the increased eddy forcing on the mean can lead
435	to an increase in the upper level zonal flow and vertical shear, as shown in Fig.8. This can
436	lead to the development of vertical velocity over the forcing region (eq.5), and with the
437	appropriate feedback of moisture supply from monsoon intraseasonal oscillations (Fig.7), the
438	condition may then be more favorable for the development of heavy rainfall episodes over
439	this region. This analysis demonstrates that the wave-mean flow interaction in the presence of

440 moisture can create favorable conditions for high-intensity precipitation events, depending on441 the potential of wave-mean interaction during the intrusion event.

442

### 5.0 Operational Model Forecast

It is clear from Fig.1 that the operational models (IITM-ERPAS, UKMO, and the 443 ECMWF have failed to capture the event. Among the three models, ECMWF performs the 444 best. It is natural to evaluate model performance in the light of the above discussion and see if 445 the relative success or failure is linked to the inability to capture the intrusion or not. The 446 verification of rainfall is plotted in Fig. 12. The plot shows that all three models 447 underestimated the rain over the Uttarakhand region. IITM-ERPAS shows low-intensity 448 rainfall over the whole of north India; the UKMO rainfall shows a high-intensity rainfall blob 449 over the Pakistan region but has very little over the Uttarakhand region. The ECMWF model, 450 however, shows some success in capturing the rainfall, but it started one day later than the 451 actual event. Thus, rainfall shows a spatial and temporal shift, which could indicate either 452 improper eddy forcing onto the mean flow resulting in improper vertical velocity or improper 453 moisture transport or a combination of both. The spatial pattern of moisture transport 454 averaged between 500-200hPa is shown in Fig. 13, which shows that the models fail to 455 forecast the low pressure-induced moisture transport. Improper moisture transport means the 456 moisture is not available for conversion to rainfall. During 15-16 June 2013, IITM-ERPAS 457 has the cyclonic circulation shifted towards the Bay of Bengal, while for UKMO, it has 458

459	shifted more towards the Arabian Sea, both missing the Indian landmass, which could cause
460	rainfall over the Pakistan region. The ECMWF model shows two anticlockwise cyclonic
461	circulations over the Arabian Sea and the Bay of Bengal during 15-16 June 2013. Finally, we
462	show $\nabla$ . $(\overrightarrow{E})$ in <b>Fig.14</b> to determine how well the models have forecasted the eddy mean
463	interaction by comparing it with the reanalysis data. The top left panel is shown for
464	reanalysis, as in <b>Fig. 10</b> . The reanalysis plots show that $\nabla$ . $(\overrightarrow{E})$ is positive (red shading) over
465	the Uttarakhand and Kashmir regions, implying acceleration of the zonal mean wind. The
466	hindcast from the models shows weaker E-vector divergence (red shades) over the
467	Uttarakhand region. Hence, the plot indicates that the eddy E-vector convergence and
468	divergence are not correctly simulated in the model forecasts over the Uttarakhand region,
469	which is also indicated by the yellow vectors (Ex, Ey). Also, the isotachs are much less dense
470	with a less wavy and intruding pattern over Kashmir, Uttarakhand, and Punjab (between 25°-
471	35°N and 70°-80°E) in the model than the reanalysis over the Indian region, implying that the
472	mean flow is also not correctly simulated in the model. Thus, the location of eddy-transient
473	forcing is also not correctly simulated in the model. Due to inappropriate eddy forcing, the
474	zonal wind height longitude profile also does not show the development of appropriate zonal
475	wind acceleration at the upper level (not shown). Inappropriate zonal wind acceleration at the
476	upper level weakened the vertical shear, thus weakening the Sutcliffe term in equation (5),
477	indicating weaker ageostrophic vertical velocity, resulting in reduced precipitation amplitude.
478	The result, therefore, suggests that the model did not capture the Uttarakhand event because

of inappropriate moisture transport from the monsoon flow and forecast of the wrong location 479 of eddy transport leading to a spatial phase shift of eddy flux divergence (north and south of 480 Uttarakhand region), resulting in improper eddy forcing over the Uttarakhand region. This is 481 also confirmed from the eddy kinetic energy (EKE) plot in Fig.11, which shows EKE plots 482 for the model forecasts. From Fig.11, it may be seen that the ECMWF model captured this 483 time sequence of EKE evolution more effectively than either the IITM-ERPAS or the UKMO 484 models. Thus, the analysis suggests that inappropriate eddy forcing and improper 485 representation of EKE in the model could be a reason for erroneous rainfall forecast in the 486 L R L L L L model. 487

488

#### 6.0 Discussion and Conclusion 489

This case study examines the life cycle of the extratropical intrusion event over 490 Uttarakhand (India) during June 2013 to understand the role of anisotropic shape, meridional 491 propagation of extratropical eddies, and their feedback onto the monsoon mean flow during 492 the 2013 Uttarakhand event. Previous studies of extreme events such as those over the 493 Uttarakhand region during June 2013 neglect the role of eddy-mean interaction in the 494 495 monsoonal region. The study reveals a potential role of underlying eddy dynamics and the inadequacy in the operational forecast models to capture the eddy dynamics. The E-vector-496 based approach that is adopted here gives a first-hand idea about the eddy forcing 497

498 mechanism, both in the observation and in the model forecast for the 2013 Uttarakhand heavy499 rainfall event.

On the synoptic scale, extratropical Rossby wave intrusion influenced the June 2013 500 heavy rainfall event over the Uttarakhand region through the southward extension of troughs 501 and associated southward shift in the subtropical jet pattern. Such synergistic evolutions are 502 documented earlier also (Raman and Rao, 1981; Kalshetti et al., 2021). At a local scale, along 503 with western Himalayan orography and moisture convergence associated with monsoon lows 504 during this time, the upper-level extratropical trough intrusion imparted a strong eddy 505 forcing, which is evident through the existence of E-vector divergence (Fig. 10) and the 506 conversion of mean kinetic energy to eddy kinetic energy (Fig.11). Thus, on 16-17 June, 507 upper-level eddy forcing accelerated the eddy circulation dynamics and developed an 508 additional shear flow that leads to the high-intensity rainfall event by amplifying the Sutcliff 509 source term (equation 5). In some locations, the mesoscale circulation can develop through 510 feedback from the large-scale flow. It can lead to cloudburst-type situations by forming a 511 super-convective system, as discussed in Houze et al. (2000). However, we have not focused 512 on this large scale to mesoscale connection in this study as we emphasized the quantification 513 of eddy forcing. The current analysis, unlike earlier studies, differentiates the eddy and mean 514 flow over the monsoonal region during this period. Our analysis only assumes the mean flow 515 and background instability associated with extratropical modes and does not assume tropical 516 depressions as a precondition. During the intrusion event, if there is background instability 517

(which can be provided by eddies of extratropical origin), wave-mean interaction can develop
ageostrophic velocity leading to high-intensity rainfall events.

The analysis presented here is based on a single extreme event. Does every extreme 520 event over the north Indian region require extratropical intrusion? It is found that high-521 intensity rainfall events over northern India (along the Himalayan belt and adjoining foothills 522 and plains) can occur in the absence of extratropical eddy forcing. Extreme events can occur 523 purely due to monsoon flow over the Indian region and are common during monsoon season 524 over different areas of India. Is there any apparent difference in the spatial pattern of rainfall 525 when an extratropical intrusion occurs? We have checked for two situations when the 526 standardized anomaly of rainfall over the north Indian region (area averaged over a box of 527 25°-40°N & 65°-90°E) is more than two standard deviations (i.e., the rainfall is high 528 intensity). In the first situation, we computed rainfall, 200-hPa wind vectors, and eddy flux 529 divergence composites in the presence of strong low frequency southward EMF transfer, and 530 in the second situation, when there is negligible EMF transfer. The low-frequency eddy 531 transfer band is defined by taking a 30-60 day filtered EMF index  $(F_L)$  area averaged over a 532 box (30°-45°N & 70°-95°E) which is then standardized. Strong southward transfer cases are 533 selected by identifying days when the standardized anomaly is less than -1. This plot is 534 shown in Fig.15. The figure compares the rainfall, wind vector at 200hPa, the divergence of 535 E vector, and E-vector components for the two extreme rainfall scenarios: with or without 536 eddy forcing. The composite is based on 24 selected cases. The plot shows that there are 537

indeed certain regions in the Himalayas and the foothills of the Himalayas, starting from 538 Uttarakhand and extending towards the east, where high-intensity rainfall occurs when there 539 is southward momentum transfer (Fig.15a). High-intensity precipitation also occurs when 540 there is negligible eddy transfer (Fig.15b). The rainfall bias plot (Fig.15c) in the last row 541 shows a positive rainfall anomaly in Uttarakhand. Although there are regional variations, 542 many locations over the foothills of the Himalayas in the eastern side of Uttarakhand show 543 positive rainfall anomalies. The wind vector plot shows that the cyclone-anticyclone pattern 544 is much closer and stronger over the north Indian region in the strong low-frequency eddy 545 transfer case. The E-vector and its divergence  $\nabla$ .  $(\vec{E})$  are shown for the two corresponding 546 scenarios in Fig.15d and Fig.15e. Strong positive values of  $\nabla$ .  $(\vec{E})$  indicate substantial 547 divergence in the low-frequency band and strong forcing on the zonal wind over the Indian 548 region (in the Kashmir region). The bias plot for divergence in Fig.15(f) also shows that these 549 extreme events associated with momentum flux transfer require strong E-vector divergence 550 above the Kashmir region. Although we have highlighted the role of EMF and its southward 551 transfer in the low-frequency mode only, all E-vector components can generate extreme 552 events. A detailed analysis for all the bands is required to understand the full implications and 553 it will be reported in a later study. Our initial results based on a case study and a composite 554 analysis in the low-frequency band suggests the usefulness of the E-vector approach in 555 understanding extratropical intrusion event. 556

The above explanation, based on the E-vector approach, was then verified in 557 operational model forecasts. Results suggest that ECMWF, ERPAS, and UKMO operational 558 models simulated upper-level circulation patterns but that the E-vector divergent field and 559 their impact on underlying atmospheric states were not correctly simulated. E-vector 560 divergence and local eddy kinetic energy are captured with varying accuracy, suggesting 561 improper eddy forcing in S2S forecast models. Also, the moisture transport in the upper level 562 is also not captured adequately (Fig.13). Thus, inappropriate eddy forcing leads to improper 563 ageostrophic and mean flow adjustment in the forecast, and inappropriate moisture transport 564 weakens moisture support, leading to a lack of skill in the model forecast. Although previous 565 analyses provided a description of the event as a Rossby wave intrusion process, the exact 566 role of Rossby wave dynamics and the eddy-mean interaction was not very clear. Our 567 analysis uncovers a series of dynamic steps to understand the reason why the rainfall is 568 underestimated in operational models during the Uttarakhand heavy rainfall event. Our study 569 also provides a diagnostic basis for evaluating the model skill using the E-vector approach 570 and can be used for model skill evaluation. 571

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580	seasonal prediction project, is a WWRP/THORPEX-WCRP joint research project established
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## Figure Caption

- Figure 1 The daily time series of rainfall from observation (TRMM) and s2s model forecasts(ERPAS, UKMO, ECMWF) area averaged over a box 29°-31°N 78°-81°E surrounding Uttarakhand region. The 95th percentile of TRMM rainfall for this box is also shown as a grey horizontal line. The ensemble spread ( $\pm$ 1SD) for the models is indicated by the shaded regions. ER-PAS and UKMO model forecasts are from 09June2013 initial conditions(ICs) and the ECMWF forecast is from 10June2013 IC.
- Figure 2 A schematic diagram describing the proposed sequence of events during the extratropical wave intrusions over the monsoonal region. Panel A shows a non-intruding situation (described as normal situaion in the figure). Panel B shows an intruding situation resulting in linear and non-linear responses, which are shown in panels C1 and C2 repectively. The resultant ageostrophic upward motion (which is forced convection in presence of monsoonal moisture background) is shown in the right side. The nonlinear effects can modulate (i.e. increase or decrease)the ageostrophic vertical velocity generation and thereby can cause flaring up of extreme rainfall events(refer text for more details).
- Figure 3 (a) The eddy momentum flux (EMF) and the eddy assymetry factor averaged over a northern box  $(30^{\circ}-40^{\circ}N;75^{\circ} 82^{\circ}E;)$  showing strong flux transfer. The eddy terms are anomaly from monthly means. Positive (negative) sign indicats anomalous northward (southward) transfer during June 2013. (b) Rainfall hovmöller plot from the 20-80 day filtered TRMM data during June 2013 averaged over longitude bands $(70^{\circ} 80^{\circ}E)$  over Indian region. (Units: EMF and Eddy Assymetry term:  $m^2 s^{-2}$  per unit area; Rainfall:  $mm \text{ day}^{-1}$ )
- Figure 4 Panels showing rainfall(shaded) patterns and vector winds at 200 hPa during 15-18June 2013. The red contours shows the zonal winds.Dates are mentioned at the top of each panels.(units: Precipitation  $mm \text{ day}^{-1}$ ; winds  $m \text{ s}^{-1}$ )

- Figure 5 Same as last figure but showing evolution of ERA5 relative vorticity pattern (shaded) and standardized eddy momentum fluxes (u'v', EMF)as contours during 15-18th June 2013. The northward (southward) transport is positive (negative). The primes and anomalies are computed with respect to the monthly mean of June 2013. The Kedarnath region of Uttarakhand state(India) which received heavy rainfall is represented as balck dots.(Units: *vorticity*  $10^{-5} \times s^{-1}$ ; *EMF*  $m^2 \, s^{-2}$ )
- Figure 6 Same as last figure but showing surface pressure anomalies from monthly mean of June 2013 and actual wind at 850 hPa. The dates are mentioned at the top of each panels. The Kedarnath region of Uttarakhand state (India) which received heavy rainfall is represented as red dots. (Units: *pressure* hPa; wind m s<sup>-1</sup>)
- Figure 7 (a)-(d) Vertically (400-200hPa) averaged moisture transport(vector) moisture divergence(shaded) during 15-18 June 2013 (dates mentioned at the top of each panel). (e)-(h) Upper level (400-200hPa) averaged static stability (Brunt Vaisalla frequency,N, anomalies from monthly mean) for the same period. The Kedarnath region of Uttarakhand state (India) which received heavy rainfall is represented as green dots.(units:div/N  $s^{-1}$ ; transport vector m $s^{-1}$ )
- Figure 8 Plots showing vertical velocity (shaded, upward positive units hPa s<sup>-1</sup>), zonal wind (black contour), relative vorticity anomaly from monthly mean(blue contour, units  $10^{-6}$  $\times$  s<sup>-1</sup>) and streamlines (magenta, units ms<sup>-1</sup>)) at 200hPa averaged for the period 15-17 June 2013. The green contour shows the potential vorticity of magnitude 1PVU.
- Figure 9 Height latitude profile evolution of vertical velocity (omega as shaded, upward positive *units*  $hPa \ s^{-1}$ ) and zonal wind (contours,*units*  $m \ s^{-1}$ )) for the days between 13-18June 2013. The plot is shown for longitudinal average range 75°E-85°E.

- Figure 10 E-Vector divergence (shaded,  $\nabla \cdot \mathbf{E}$ ), E-Vectors (yellow) for the period averaged between 10-30June 2013. The zonal wind contours for the days averaged between 14-18 June 2013 are also plotted to show the average jet position during the extreme event days. The eddies are defined by taking average for the period 10-30 June 2013.
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- Figure 12 A comparison of rainfall evolution (shaded) for the period 15-18 June 2013 for different operational models and TRMM observation data. (unit: mm  $day^{-1}$ )
- Figure 13 Plot showing the comparison of mositure transport vectors and moisture divergence (shaded) based on ERA5 reanalysis and different s2s model runs (mentioned in top of each panels).(units: div  $s^{-1}$ ; transport vector m  $s^{-1}$ )
- Figure 14 Plot showing the verification of E-vectors and E-vector divergence (shaded,  $\nabla \cdot \mathbf{E}$ ) for different model runs with respect to observation for the days averaged between 14-18 June 2013.
- Figure 15 Rainfall and wind anomaly composites when (a) the standardized rainfall anomaly over north-Indian region is more than 2 standard deviation and the standardized EMF index area-averaged over a northern box (30°-45°N, 70°-95°E) is more than 2 standard deviation, (b) same as (a) but when the standardized EMF is close to zero. (c) shows the rainfall and wind vector bias for these two cases i.e. (a)-(b). (d) E-vector divergence composite and plot of E-vector at 200hPa for the same criteria as (a).(e) is same as (b) but for E-vector divergence composite and plot of E-vector at 200hPa. (f) same as (c) but showing the bias in E-vector divergence and E-vectors.



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1	-Dynamics of Tropical-Extratropical Interactions in the S2S
2	forecast models and observations during the Monsoon Season
3	<del>over the Himalayan regions</del>
4	Eddy transport, Wave-mean flow interaction, and Eddy forcing
5	during the 2013 Uttarakhand Extreme Event in the Reanalysis
6	and S2S Retrospective Forecast Data
7	Mahesh Kalshetti <sup>1,3</sup> , Rajib Chattopadhyay <sup>1,2</sup> -, Kieran <u>M R</u> Hunt <sup>4</sup> , R
8	phani <sup>4</sup> Phani <sup>1</sup> , Susmitha Joseph <sup>1</sup> , DR Pattanaik <sup>2</sup> , AK Sahai <sup>1</sup>
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20	
21	Keywords
22 23	Extreme events; Extratropical-Tropical teleconnection; Indian monsoon, subseasonal forecasts
24	
25	Abstract
26	In this study, to explore the wave-mean interaction during the monsoon season, we
27	investigate (a) the potential role of transient eddy forcing and the wave-mean interaction on the
28	monsoon weather during the June 2013 Uttarakhand (India) extreme <u>heavy</u> rainfall event over
29	the Himalayan regions (especially the Uttarakhand State of India and the nearby regions) and,
30	(b) how they are captured in a set of operational models. Some studies have pointed out how
31	prolonged breaks can occur due to extratropical trough incursions. However, there is a lack of
32	clarity on how transient eddy forcing associated with such interactions can lead to modulation
33	of monsoonal circulation, leading or whether such interaction can lead to heavy rainfall events.
34	E-vector fields are analyzed to quantify about the eddy forcing from the extratropical
35	transient eddies and the feedback mechanism between the transient eddies and the mean flow
36	during June 2013. Analysis reveals that along with local factors (orography, moisture
37	convergence), the large-scale heavy rainfall event over the Uttarakhand region during 16-17
38	June 2013 event was influenced by eddy forcing due to the intrusion of extratropical Rossby

waves over the Indian region. The location of eddy affects the location of regional occurrence 39 40 of the eddy-mean interaction. Model hindcast analysis results suggest that operational models cannot forecast the upper-level eddy forced circulation patterns, and the improper 41 representation of the E-vector divergence field leads to the underestimation of intensity and the 42 spatial pattern of rainfall-pattern. 43 44 45 **Introduction** 46 47 It is well known that extratropical-tropical (E2T) interaction can cause prolonged breaks (low rainfall spells) over the Indian region due to intrusion of extratropical troughs and ridges 48 during the monsoon season (RAMASWAMY, 1962; Raman and Rao, 1981; Krishnan et al., 49 2009; Fadnavis and Chattopadhyay, 2017). However, in recent years studies have shown that 50 extreme rainfall events over the Himalayan region are sometimes associated with the 51 52 coexistence of both low-frequency monsoon intraseasonal oscillations and extratropical eddies (or Rossby waves or Western Disturbances). The eddy transport and fluctuations are evident 53 through the associated heat and momentum transport towards or out of the tropics during the 54 monsoon time, as shown in Kalshetti et al., (2020). Disturbances originating in the extratropics 55 can locally shift the jet stream equatorward and thus disturbs the synoptic setup of the Indian 56 summer monsoon. This intrusion is supposedly linked to extreme rainfall events like the 2010 57

58	Pakistan and 2013 Uttarakhand flood events (Hong et al., 2011; Lau and Kim, 2011; Joseph et
59	al., 2015; Vellore et al., 2016a; Sooraj et al., 2020) and many similar weather events.
60	How does the extratropical intrusion create an extreme event? An extreme event can be
61	dynamically defined as zones of strong vertical velocities with ample moisture supply from the
62	boundary layer or lower troposphere (such as during monsoon time), which is likely due to
63	strong ageostrophic components (Bohlinger et al., 2019). Classic numerical experiments such
64	as from (Hoskins and Karoly, 1981) and some recent case studies based on Eurasian blocking
65	suggests that 2010 western Russian heatwave is forced by downstream Rossby wave
66	propagation, with high-frequency Rossby waves also propagate southward. Thus, Rossby
67	waves are often precursor of extreme weather events. The leading trough of the Rossby wave
68	train triggered upward motion to the east, favoring deep convection leading to extreme flooding
69	over Pakistan during 2010(Hong et al., 2011; Lau and Kim, 2011). Studies have identified the
70	thermal forcing and background moist processes over Indo-Pakistan arid region in the 16-18
71	days prior and are shown to be precursors for extreme rainfall events over the monsoon core
72	zone (Sooraj et al., 2020).

Similarly, some other studies suggest extreme rainfall events such as those that occurred
over the Uttarakhand region during June 2013, result from an occluded frontal system
developed in response to western disturbance and monsoon low. Occluded fronts caused
mechanical lifting, thus the formation of the large organized storm over the study region

77	(Chevuturi and Dimri, 2016). The validity of the frontal theory of ageostrophic upliftment
78	requires the validity of the ratio $NH/f$ (N is Brunt-Väisälä frequency, $f$ is Coriolis parameter and
79	H being the depth of the fluid) as described in Hoskins (1982). For tropical atmosphere, the
80	validity of this ratio is not well defined due to smallness of f. Also, as (Charney, 1969)
81	suggested, such large scale lifting in the stably stratified tropical climate is not easily (or
82	generally) possible due to lack of vertical coupling unless it occurs in the regions of active
83	tropical deep cumulus convection on a much smaller spatial scale than the Rossby radius. Since
84	the destabilization by latent heating due to moisture supply may be less over the orographic
85	region over the Himalayas (where the event occurred) than the transient eddy forcing on the
86	mean background (Wills and Schneider, 2018), it is possible that alternate eddy-driven theories
87	may be more helpful.
87 88	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as
87 88 89	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably
87 88 89 90	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably during monsoon time or, more specifically, due to the presence of low-pressure systems over
87 88 89 90 91	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably during monsoon time or, more specifically, due to the presence of low-pressure systems over the Indian subcontinent. Vellore et al. (2016b)and Hunt et al (2018a, 2018b) have looked into
87 88 89 90 91 92	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably during monsoon time or, more specifically, due to the presence of low-pressure systems over the Indian subcontinent. Vellore et al. (2016b)and Hunt et al (2018a, 2018b) have looked into the extratropical intrusions of western disturbances in a detailed manner and described the
87 88 89 90 91 92 93	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably during monsoon time or, more specifically, due to the presence of low-pressure systems over the Indian subcontinent. Vellore et al. (2016b)and Hunt et al (2018a, 2018b) have looked into the extratropical intrusions of western disturbances in a detailed manner and described the presence of monsoon system and western disturbances as a sufficient condition for extreme
87 88 90 91 92 93 94	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably during monsoon time or, more specifically, due to the presence of low-pressure systems over the Indian subcontinent. Vellore et al. (2016b)and Hunt et al (2018a, 2018b) have looked into the extratropical intrusions of western disturbances in a detailed manner and described the presence of monsoon system and western disturbances as a sufficient condition for extreme events indicating presence of double eddies (western disturbances and monsoonal low pressure
87 88 90 91 92 93 94 95	may be more helpful. Several studies have explored the eddy (i.e., transient synoptic disturbance) perspectives as mentioned earlier. The observation suggests that such extreme rainfall events occur invariably during monsoon time or, more specifically, due to the presence of low-pressure systems over the Indian subcontinent. Vellore et al. (2016b)and Hunt et al (2018a, 2018b) have looked into the extratropical intrusions of western disturbances in a detailed manner and described the presence of monsoon system and western disturbances as a sufficient condition for extreme events indicating presence of double eddies (western disturbances and monsoonal low pressure systems) during which interaction occurs as a plausible theory. Hunt et al (2018a, 2018b), based

9	classes of such events with highlights on such presence of double eddies(western disturbances
9	and monsoonal high frequency disturbances or low pressure systems). Another recent study
9	(Hunt et al., 2021) explained the extreme events in terms of the coexistence of the monsoon
10	system and western disturbances during the extreme event for different classes of situations.
10	They describe that extreme precipitation occurs as a result of their interactions, but their
10	analysis uses the perspective of Hanley et al. (2001), which assumes pre-existence of an intense
10	cyclonic storm (with substantial azimuthal eddy flux transfer, EFCs) co-located with jet stream
10	and that this storm interacts with the upper-level troughs when EFCs cross certain thresholds.
10	In such situations, the vertical coupling would be strong, as shown by Charney (1969) and
10	developments of extreme events are possible. This EFC-based criterion is, however, not
10	quantitatively verified in these studies.
10	The studies described above
10	
11	<u>1. Introduction</u>
11	It is well known that extratropical-tropical (E2T) interaction can cause prolonged breaks
11	(low rainfall spells) over the Indian region due to intrusion of extratropical troughs and ridges
11	during the monsoon season (Ramaswamy, 1962; Raman and Rao, 1981; Krishnan et al., 2009;
11	Fadnavis and Chattopadhyay, 2017). However, in recent years, studies have shown that high-
11	intensity rainfall events over the Himalayan region are sometimes associated with the

116	coexistence of low-frequency monsoon intraseasonal oscillations and extratropical eddies (or
117	Rossby waves or Western Disturbances). The eddy transport and fluctuations are evident
118	through the associated heat and momentum transport towards or out of the tropics during the
119	summer monsoon, as shown in Kalshetti et al. (2020). Disturbances originating in the
120	extratropics can locally shift the jet stream equatorward and thus disturb the synoptic setup of
121	the Indian summer monsoon. These intrusions are supposedly linked to high-intensity rainfall
122	events like the 2010 Pakistan and 2013 Uttarakhand floods (Hong et al., 2011; Lau and Kim,
123	2011; Joseph et al., 2015; Vellore et al., 2016a; Sooraj et al., 2020) and many similar weather
124	events.
125	How does the extratropical wave intrusion create high-intensity rainfall over Northern
126	India, especially over the Himalayan region? The high-intensity rainfall event can be
127	dynamically defined as zones of strong vertical velocities with ample moisture supply from the
128	boundary layer or lower troposphere (such as during monsoon time), which is likely due to
129	strong ageostrophic components (Bohlinger et al., 2019). Can a Rossby wave intrusion cause
130	this high-intensity rainfall or other types of extreme events, e.g., heatwaves? Classic numerical
131	experiments such as from Hoskins and Karoly (1981) and some recent case studies based on
132	Eurasian blocking suggests that the 2010 western Russian heatwave was forced by downstream
133	Rossby wave propagation (Trenberth and Fasullo, 2012), with high-frequency Rossby waves
134	also propagated southward during the end of July (Lau and Kim, 2011). Thus, Rossby waves
135	are often a precursor of extreme heatwave events. Similarly, the leading trough of the Rossby

136	wave train triggered upward motion to the east, favoring deep convection leading to extreme
137	rainfall event-related flooding over Pakistan during 2010 (Hong et al., 2011; Lau and Kim,
138	2011). Likewise, extratropical Rossby wave-linked convection is also reported in the tropical
139	Pacific (Kiladis, 1998). Heavy rainfall, especially extreme events, is associated with mesoscale
140	convective systems. Several studies on midlatitude convection showed that midlatitude Rossby
141	waves are reinforced by mesoscale systems (Lillo and Parsons, 2017; Parsons et al., 2019),
142	which create downstream forced response. Some studies have also shown how mesoscale
143	convective systems over the Pacific warm pool region get reinforced by Rossby-kelvin
144	waves(Houze et al., 2000), indicating the role of large-scale waves in local mesoscale
145	precipitating systems. In addition to waves, studies have also identified thermal forcing and
146	background moist processes over the arid region of Indo-Pakistan in the 16 to18 days prior and
147	have shown them to be precursors for extreme rainfall events over the monsoon core zone
148	(Sooraj <i>et al.</i> , 2020).
149	As mentioned in the last paragraph, the state of Uttarakhand in India witnessed heavy
150	rainfall during 15-17 June 2013 with widespread reporting of a "cloudburst"-like scenario over
151	the Kedarnath region in Uttarakhand. Some studies have demonstrated the role of mesoscale
152	convective systems in cloudburst situations(Parida et al., 2017). Other studies suggest high-
153	intensity rainfall events such as those that occurred over the Uttarakhand region during June
154	2013 result from an occluded frontal system that developed in response to a western
155	disturbance and monsoon low (Chevuturi and Dimri, 2016). The validity of the frontal theory
	8

156	of ageostrophic uplift requires the validity (i.e., finiteness) of the ratio NH/f(N is Brunt-Väisälä
157	frequency, $f$ is the Coriolis parameter, and $H$ is the depth of the fluid) as described in Hoskins
158	(1982). In the tropics, the validity of this ratio is not well defined due to the smallness of $f$ .
159	Also, as (Charney 1969) suggested, such large-scale lifting in the stably stratified tropical
160	climate is not easily (or generally) possible due to lack of vertical coupling unless it occurs in
161	the regions of active tropical deep cumulus convection.
162	In addition to such localized extreme events during 16-17 June, heavy to very heavy rainfall
163	occurred during this time over larger regions (in the synoptic scale, e.g., refer fig.2 of Joseph
164	et al., (2015) or Fig.12). The coexistence of synoptic-scale heavy rainfall and Rossby wave
165	intrusion may indicate that the transient eddies may be a dominant factor in comparison with
166	other factors like local destabilization due to latent heating, which is already reported (e.g., Fig.
167	9b of (Wills and Schneider, 2018).
168	This eddy (i.e., transient synoptic disturbance) perspective has also been explored in several
169	studies. Vellore et al. (2016b) and Hunt et al. (2018a, 2018b), using cluster composite analysis,
170	looked into several extratropical intrusions of western disturbances in a detailed manner and
171	described the presence of monsoon low-pressure systems and western disturbance as a
172	necessary condition for heavy events. Another recent study (Hunt et al., 2021) explained the
173	extreme events in terms of the coexistence of monsoon low-pressure systems and western
174	disturbances. They found that high-intensity precipitation occurs as a result of their

175 interactions, but their analysis uses the perspective of Hanley et al. (2001), which assumes the pre-existence of an intense cyclonic storm (with substantial azimuthal eddy flux transfer, EFCs 176 which is a measure of local eddy forcing) co-located with the jet stream and that this storm 177 interacts with the upper-level troughs when the eddy-induced azimuthal flux transfer measured 178 by EFCs cross certain thresholds and provide necessary eddy forcing. This EFC-based criterion 179 is, however, not quantitatively verified in these studies. 180 The studies described above indicate that the orography, monsoonal moisture flow, 181 stability, and waves/eddies can contribute to the extreme as well as heavy rainfall event, but do 182 not go into explaining this monsoon midlatitude waveeddy interaction scenario as an eddy 183 forcing problem on the mean flow (i.e., eddy-mean interaction perspective), and hence, though 184 the earlier analysis highlights exchanges, they neglect a wave-mean interaction perspective, 185 i.e., interaction terms in vorticity tendency equations obtained by partitioning the fields into 186 the mean and eddy terms(e.g. cftransient wave forcing perspective (e.g., refer eq. 1). in sec.3). 187 In the current analysis, we will focus on how (or whether) wave-mean interaction can be used 188 189 to explain ageostrophic vertical velocities and rainfall that occurs during such events. Unlike earlier studies, we will explain the intrusions in terms of transient eddy forcing (such as eddy 190 flux convergence or divergence) associated with the Rossby waves intruding over the Indian 191

192 region because of large-scale precursors, e.g., circumglobal teleconnection. Such impacts of

193 high and low-frequency eddies (Synoptic Eddy Low-Frequency or SELF feedback) are studied

194 forSuch interactions perspective are often useful to study the mid-latitude low-frequency
195 oscillations (Jin, 2010).

196	The eddy forcing approach is particularly useful if we want to know why the dynamical
197	models fail to forecast such events at enough lead times for issuing alerts(i.e., whether there is
198	inappropriate eddy forcing in the model), especially in the weather to extended-range forecast
199	time scale. If such events originate due to low frequency (e.g., circumglobal) teleconnection
200	pattern, which is supposed to have better predictability (because they are slowly evolving), why
201	are such events not captured with good fidelity in the dynamical model forecast (e.g., refer
202	Fig.1) as they are captured in midlatitude synoptic-scale forecasts? Fig.1 shows that several
203	operational dynamical models lack skill in predicting the event even at a one-week lead-time.
204	The Uttarakhand extreme event of 2013 is popularly studied by many researchers and is a well-
205	known event in which tropical-extratropical interaction is well documented (Dube et al., 2014;
206	Vellore and Jayant, 2014; Joseph et al., 2015; Pattanaik et al., 2015; Singh and Chand, 2015;
207	Vellore et al., 2016a; Kaur and Gupta, 2017). The question that arises here is that how to
208	explain lack of skill in a dynamical framework. We hypothesize that eddy-mean interaction
209	framework can be useful in understanding the skill (or lack of it) in dynamical models
210	In the current paper, we focus on the 2013 extratropical intrusion event over Uttarakhand
211	(India). We elaborate on two aspects: (a) whether wave-mean interaction was significant during
212	this event in addition to the earlier linear theories used to explore such events and (b) based on
213	this transient eddy wave mean interaction formulation, whether such steps can be retraced or

214	evaluated in the dynamical models to understand the lack of skills in these models in
215	forecasting these events. We computed the transient eddy forcing during the extreme event
216	using the <i>E-vector</i> approach (Andrews and McIntyre, 1976; Hoskins et al., 1983), which
217	explains the eddy dynamical pathway for such events near subtropical jet streams (Sec.3).
218	Models fail to predict the rainfall pattern and amplitude over Uttarakhand with sufficient
219	lead-time Fig.1 shows that several operational dynamical models lack skill in predicting the
220	rainfall event over Uttarakhand even at a one-week lead-time. Several models have these
221	problems with events related to Rossby waves (e.g., forecast bursts as referred in (Lillo and
222	Parsons, 2017). In addition to understanding the eddy-mean interaction during the intrusion,
223	the eddy forcing approach can be useful if we want to know why the dynamical models fail to
224	forecast such events at sufficient lead times for issuing alerts, especially in the weather to the
225	extended-range forecast time scale. We hypothesize that the eddy-mean interaction perspective
226	can be helpful in understanding the skill (or lack of it) in dynamical models.
227	The Uttarakhand extreme event of 2013 and the high-intensity rainfall over large regions
228	has been studied by many researchers and is a well-known event in which tropical-extratropical
229	interaction is well documented (Dube et al., 2014; Vellore and Jayant, 2014; Joseph et al.,
230	2015; Pattanaik et al., 2015; Singh and Chand, 2015; Vellore et al., 2016a; Kaur and Gupta,
231	2017). Hence, owing to the good documentation of linear perspective, the current paper selects
232	this event and discusses the eddy-mean interaction perspective. We elaborate on two aspects:
233	(a) whether wave-mean interaction was significant during this event in addition to the earlier
	12

234 linear theories used to explore such events and (b) based on this transient eddy wave mean interaction formulation, whether such steps can be retraced or evaluated in the dynamical 235 models to understand the lack of skill in these models when forecasting such events. The study 236 is arranged as follows: Sec.2 describes the study area, data used, and formal methodology. In 237 sec.3, we computed the transient eddy forcing during the extreme event using the *E-vector* 238 239 approach (Andrews and McIntyre, 1976; Hoskins et al., 1983), which explains the eddy dynamical pathway for such events near subtropical jet streams. Sec.4 describes the application 240 of the E-vector approach while describing the life cycle of tropical-extratropical interaction 241 while describing the intrusion event during the Uttarakhand heavy rainfall event (June 2013). 242 Sec.5 describes the operational forecasts of the event, and the results are discussed, 243 summarized, and concluded in Sec.6. 244 2. Study area, Data, and Methods 245 2.1. Study area 246 247 The natural disaster in the form of landslide and flash flood occurred due to heavy precipitation over Uttarakhand region (29°-31°N, 78°-81°E), India on 16-17 June 2013. The 248

249 Uttarakhand is part of the region of complex orography along Western Himalaya. The Western

250 Himalaya region surrounds snow-covered peaks, crest, glaciers, valleys, and perennial river

251 basins (Parida *et al.*, 2017).

## 252 2.2. Observation Data

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253 The atmospheric (retrospective) reanalysis datasets from the fifth-generation European Centre for Medium-Range Weather Forecasts (ECMWF) Atmospheric Reanalysis of the global 254 climate (ERA5) is used as a proxy for observation of dynamical variables. ERA5 data is 255 produced by Copernicus Climate Change Service (C3S) at ECMWF. ERA5 climate data is 256 available in a  $0.25^{\circ} \times 0.25^{\circ}$  grid from 1979 to within five days of real-time (Hersbach *et al.*, 257 258 2020). The primary data used in this study are zonal (u ms<sup>-4</sup>), meridional (v ms<sup>-4</sup>), vertical (w Pa s<sup>-1</sup>) wind, temperature (K) at 200hPa, and geopotential height (m) at 500hPa. The observed 259 precipitation is obtained from Tropical Rainfall Measuring Mission (TRMM) Multi-satellite 260 Precipitation Analysis (TMPA, 3B42) (Huffman et al., 2007). 261

262 **2.3.** *Models* 

The IITM-ERPAS or ERPAS (Extended Range Prediction for Application to Society, run 263 at Indian Institute of Tropical Meteorology (IITM), Pune, India and is also the current 264 operational extended range model of India meteorological Department), UKMO (United 265 Kingdom Meteorological Office, Met Office, Devon, United Kingdom), and ECMWF (European 266 Centre for Medium-range Weather Forecasts, ECMWF, Reading, United Kingdom) are the S2S 267 reforecast models used in this study. All these S2S models are used for real-time operational 268 forecasts. All the observed and model data set regridded to 1° ×1° spatial resolution. Except 269 for IITM-ERPAS (Abhilash et al., 2014) the UKMO, and ECMWF forecast data access through 270 the S2S database maintained by the ECMWF (Vitart et al., 2017). ERPAS data is obtained 271
272	from IITM-IMD operational database maintained at IITM. Each S2S model has a control
273	member (using a single unperturbed initial condition) and several perturbed members produced
274	for sampling uncertainty in the initial condition. Table 1 shows the basic features of S2S
275	models (Vitart et al., 2017). In the present study, the model forecast is taken from the closest
276	available initial condition on 9 <sup>th</sup> June 2013 for ERPAS and UKMO and for 10 <sup>th</sup> June 2013 for
277	ECMWF.
278	The natural disaster, in the form of landslides and flash flooding, occurred due to heavy
279	precipitation over the Uttarakhand region (29°-31°N, 78°-81°E), India on 16-17 June 2013.
280	Uttarakhand is part of the region of complex orography along the Western Himalayas. The
281	Western Himalaya region surrounds snow-covered peaks, crest, glaciers, valleys, and perennial
282	river basins (Parida et al., 2017).
283	2.2. Observation Data
284	The atmospheric (retrospective) reanalysis dataset from the fifth-generation European
285	Centre for Medium-Range Weather Forecasts (ECMWF) Atmospheric Reanalysis of the global
286	climate (ERA5) is used as a proxy for observation of dynamical variables. ERA5 data are
287	produced by the Copernicus Climate Change Service (C3S) at ECMWF. ERA5 climate data is
288	available in a $0.25^{\circ} \times 0.25^{\circ}$ grid from 1979 to within five days of real-time (Hersbach <i>et al.</i> ,
289	2020). The primary data used in this study are zonal ( $u$ , ms <sup>-1</sup> ), meridional ( $v$ , ms <sup>-1</sup> ), vertical ( $\omega$ )
290	Pa s <sup>-1</sup> ) wind, temperature (K) at 200hPa, and geopotential height (m) at 500hPa. The observed
•	

291	precipitation is obtained from Tropical Rainfall Measuring Mission (TRMM) Multi-satellite
292	Precipitation Analysis (TMPA, 3B42) (Huffman et al., 2007). 3B42 data contains a gridded,
293	satellite-based merged infrared precipitation data (mm/hr), with a 3-hour temporal resolution
294	and a 0.25-degree spatial resolution. We have converted the sub-daily data to daily data.
295	<u>2.3. Models</u>
296	The Indian Institute of Tropical Meteorology (IITM) extended range forecast (IITM-
297	ERPAS) runs (Abhilash et al., 2014; Sahai et al., 2019), UKMO (United Kingdom Met Office),
298	and ECMWF (European Centre for Medium-range Weather Forecasts) forecast runs are the
299	S2S reforecast model runs used in this study. All the observed and model data set regridded to
300	$1^{\circ} \times 1^{\circ}$ spatial resolution and daily temporal resolution with appropriate pre-processing. Except
301	for IITM-ERPAS (Abhilash et al., 2014), the UKMO and ECMWF forecast data were accessed
302	through the S2S "instantaneous and accumulated" database maintained by ECMWF (Vitart et
303	al., 2017). ERPAS data is obtained from the IITM-IMD operational database maintained at
304	IITM. Each S2S model has a control member (using unperturbed initial conditions) and several
305	perturbed members produced for sampling uncertainty in the initial conditions. In the present
306	study, the model forecast is taken from the closest available initial condition on 9th June 2013
307	for ERPAS and UKMO and for 10th June 2013 for ECMWF when the forecast of 16-17 June
308	lies in the synoptic range. Spatial plots are shown for the ensemble mean (IITM-ERPAS-16-

309 <u>member mean; UKMO-7 member (including one control member); ECMWF-11 member</u>
 310 (including one control member) unless stated otherwise.

#### 311 *3.0 A schematic description of the Life Cyclelife cycle of extratropical intrusion*

The extratropical eddies have are associated with colder extratropical air massmasses and 312 313 strongstronger westerly momentum atin the upper level compared tolevels than tropical monsoonal flow-at upper level. The convergencetransport of this eddy momentum and heat 314 flux over the Indian region in the presence of low-frequency monsoonal background (monsoon 315 intraseasonal oscillations and low-pressure systems (Goswami, 2012)(Goswami, 2012)) can 316 significantly impact- the local flow as it imparts additional heat and momentum in the local 317 budget terms (cf. eq. one and eq.2 in sec.3.2). The event life cycle can thus can be hypothesized 318 to have three major stages: (i) southward *digging* (or intrusion <del>)</del>with amplification) of troughs 319 associated with <u>a</u> meandering of <del>Jetstream, (the subtropical jet stream, (ii)</del> transient eddy forcing 320 onto the mean flow, and (iii) reversal of jet stream (or weakening of southward intrusion) 321 position with the restoration of monsoon flow. These are described below. 322

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# 3.1 Linear Process: Intensification of digging trough over the Indian Region during the monsoon season

Rossby <u>wavewaves</u> show meridional propagation, but the southward intrusion of the Rossby wave <u>is prohibited toin</u> the <u>deep</u> tropical region <u>into the Indian regionis generally restricted</u> due to the presence of critical latitudes where the phase velocity of the wave approaches the

328	zonal (westerly) wind speed, below which no zonal propagation of Rossby wave phases is
329	seen.and the waves are absorbed along the critical latitudes as it propagates equatorward
330	(Karoly and Hoskins, 1982). Normally Rossby waves break, and mixing of potential vorticity
331	(PV) occurs with the decay of the wave- <u>leading to mid-latitude to tropical exchange</u> (Homeyer
332	and Bowman, 2013). However, Rossby waves can dig deep south into the tropical region
333	provided the critical latitudes shifted southward due to due to the ducting effect or southward
334	extension of westerlies over a certain region (Webster and Holton, 1982)(Webster and Holton,
335	<u>1982</u> ). <u>Southward The southward shifting/meandering of the westerly jet stream southward due</u>
336	to the presence of the blocking high can provide such ducts where Rossby wavewaves can
337	propagate into the tropics. At the same time, the cyclonic Rossby wave breaking can be
338	prevented, and it can intensify over the Indian region in the presence of monsoonal instabilities.
339	It can be seen as follows:

340 Consider the conservation of 
$$PV \div \frac{\zeta + f}{\frac{\partial p}{\partial \theta}} \approx Const in the absence of diabatic heating: \frac{\zeta \pm f}{\frac{\partial p}{\partial \theta}}$$
  
341  $\approx Const$ 

If a cyclonic phase Rossby wave (positive vorticity) comes to into a region of higher thickness  

$$(\frac{\partial p}{\partial \theta}, (\frac{\partial p}{\partial \theta}), (\frac{\partial p}{\partial \theta}))$$
 and *f* decreases, relative vorticity has to increase. For more detailed explanations of  
the PV aspects of synoptic developments, the reader refers to case studies discussed in  
(Hoskins, 1997). Hoskins (1997). Thus, the cyclonic vorticity induced by the Rossby wave can  
18

346 intensify (or <u>atleast at least its</u> amplitude does not decay) depending on the strength of the denominator, as a southward extension means a decrease in f. Typically such thickness would 347 increase in the presence of strong heating. Hence the Rossby wave would strengthen and not 348 break or weaken in the presence of background instability that modifies  $\frac{\partial p}{\partial A}$ . Such increased 349 background thickness during some period can be provided by latent heat release associated 350 with the intraseasonal monsoon oscillation modes of subseasonal variability when the organized 351 convection stays over the foothills of the Himalayas. 352 The intruded (or possibly intensified) Rossby wave can create a digging trough, leading 353 to indirect ageostrophic circulation with the high vertical velocity at the jet exit region. The 354 upper-level divergence force lower-level convergence via the indirect circulation (e.g., Chapter 355 2.3 and if moisture present, it support convection. This forced convection is then enhanced due 356 to indirect circulation. 357 Fig. 2.7 of Lackmann (2012). In addition to such direct effects, the 358 359 strengthening or weakening of the westerly or the Rossby wave depends on the growth of the eddy instability, which can extract energy from the mean flow- (Held and Phillips, 1989). The 360 response of the mean state due to the transient eddy forcing would beis essential to control the 361 growth or decay- of the Rossby wave. This nonlinear process is discussed next. 362 363 3.2 Nonlinear process: the transient eddy forcing of the mean flow and 364 **Monsoonal** Monsoon time mean background 19

365	There are numerous studies on empirical evidence of feedback of transient eddies to
366	maintain time mean flow and how eddies extract energy from the mean flow or give energy to
367	the mean flow(Lau and Holopainen, 1984; G. Branstator, 1995; Jin, 2010; Tan et al., 2014). In
368	particular recent study shows transient eddy feedback for Rossby waves over the orographic
369	region is more important than latent heating (Wills and Schneider, 2018). The eddy mean
370	interaction for the horizontal momentum equations assuming zonal wind (u) and meridional
371	wind (v) compartmentalized into mean and transient eddy components ( $\mathcal{V} = \overline{\mathcal{V}} + \mathcal{V}', \mathcal{V} = (u,v)$
372	) ( Hoskins et al. 1983; James 1994; Williams et al. 2007) is given by:
373	$\frac{\overline{D}\overline{u}}{Dt} = f v_{am} + \nabla . E $ (1a)
374	$\frac{\overline{D}\overline{v}}{Dt} = -fu_{am} - (\overline{u'v'})_x - (1b)$
375	$\frac{\overline{D}}{Dt} = \overline{u}\frac{\partial}{\partial x} + \overline{v}\frac{\partial}{\partial y} - ()_x = \frac{\partial}{\partial x} + \overline{v}$ is There are numerous studies demonstrating empirical
376	evidence of the role of the transient eddies in maintaining the time-mean flow and how eddies
377	extract energy from the mean flow or give energy to the mean flow (Lau and Holopainen, 1984;
378	G. Branstator, 1995; Jin, 2010; Tan et al., 2014). In particular, a recent study (Wills and
379	Schneider, 2018) showed transient eddy feedback for Rossby waves over the orographic region
380	is more important than latent heating. The eddy mean interaction for the horizontal momentum
381	equations, assuming zonal wind $(u)$ and meridional wind $(v)$ compartmentalized into mean and
382	transient eddy components: $(\underline{\mathcal{V}} \equiv \overline{\underline{\mathcal{V}}} + \underline{\mathcal{V}}', \underline{\mathcal{V}} \equiv (u,v))$ (Hoskin et al. 1983; James 1994;
383	Williams et al. 2007) is given by:

$$\frac{B}{Dt} = f \pm \nabla_{...}E - ....(1a)$$

$$\frac{D}{Dt} = -fu_{am} = (\overline{u'v'})_x - ....(1b)$$

$$\frac{D}{Dt} = \overline{u_{dx}} \pm \overline{u}_{dy}^{\theta} + (\nabla_{x} + \overline{v})_{x} - ....(1b)$$

$$\frac{D}{Dt} = \overline{u_{dx}} \pm \overline{u}_{dy}^{\theta} + (\nabla_{x} + \overline{v})_{x} - ....(1b)$$

$$\frac{D}{Dt} = \overline{u_{dx}} \pm \overline{u}_{dy}^{\theta} + (\nabla_{x} + \overline{v})_{x} - ....(1b)$$

$$\frac{D}{Dt} = \overline{u_{dx}} \pm \overline{u}_{dy}^{\theta} + (\nabla_{x} - \nabla_{x} + \overline{v})_{x} + ....(1b)$$

$$\frac{D}{Dt} = \overline{u_{dx}} \pm \overline{u}_{dy}^{\theta} + (\nabla_{x} - \nabla_{x} + \overline{v})_{x} + ....(1b)$$

$$\frac{D}{Dt} = \overline{u_{dx}} \pm \overline{u}_{dy}^{\theta} + (\nabla_{x} - \nabla_{x} + \overline{v})_{x} + ....(1b)$$

$$\frac{D}{Dt} = \overline{v_{dx}} \pm \overline{v_{dy}} + ....(1b)$$

$$\frac{D}{Dt} = \overline{v_{dx}} \pm \overline{v_{dy}} + ....(1b)$$

$$\frac{D}{Dt} = .....(1b)$$

$$\frac{D}{Dt} = ....(1b)$$

$$\frac{D}{Dt} = ....(1b)$$

$$\frac{D$$

I

401	other factors like friction are negligible in the upper level), i.e., eddies can force the mean
402	flow. Divergence of the E-vector can cause acceleration of mean flow, and convergence can
403	decelerate the mean flow. In a monsoonal setting, if we consider <u>take</u> the terms in the LHS of
404	Eq.1 as representing the mean monsoonal quasi-stationary background, the second term in the
405	RHS can be interpreted as the eddy forcing term arising due to extratropical intrusions or
406	similar low-pressure systems. Depending on the sign of the eddy convergence, the eddy can
407	either grow (or extract energy from mean flow) or they can dissipate by supplying energy to
408	the monsoonal mean flow. To understand and estimate the impact of this eddy forcing, we will
409	use the (also refer to (Held and Phillips, 1989). To understand and estimate the impact of this
410	eddy forcing during the Uttarakhand event, we will use this E-vector formalism.
411	The E-vector is also related to transient eddy vorticity flux convergence and divergence. It can
412	be shown that, $\nabla \cdot \overline{u'\zeta'} \approx \frac{d}{dy} (\nabla \cdot E)$ (3)
413	Then from Eq.1 and Eq.3 we can see that an E-vector framework provides a first-hand
414	idea of the transient eddy forcing and zonal asymmetries imparted by the transient eddies
415	(Hoskins et al., 1983; Trenberth, 1986)in the momentum and the vorticity equations. Also
416	plotting of E-vectors, i.e. (Ex,Ey) can give an idea about the direction and location of eddy
417	convergence. The E-vector approach can be used to study both zonally asymmetric tropical
418	(Leroux et al., 2010) and extratropical transport patterns (Novak et al., 2015).

# The E-vector is also related to transient eddy vorticity flux convergence and divergence. It can 419 $\underline{\nabla. \overline{u'\zeta'}} \approx \frac{\underline{d}}{dy}(\underline{\nabla.E})....(3)$ be shown (Hoskins et al., 1983) that: 420 Then, from Eq.1 and Eq.3, we can see that an E-vector framework provides a first-hand 421 idea of the transient eddy forcing and zonal asymmetries imparted by the transient eddies 422 423 (Hoskins et al., 1983; Trenberth, 1986) in the momentum and the vorticity equations. Also, plotting of E-vectors, i.e. (Ex, Ey), can give an idea about the direction and location of eddy 424 convergence. The E-vector approach can be used to study both zonally asymmetric tropical 425 (Leroux et al., 2010) and extratropical transport patterns (Novak et al., 2015). The above 426 discussion is only based on the mechanical effects of eddies. A complete quasigeostrophic 427 discussion including the thermal effects of eddies by defining a third component in E-vector 428 using temperature (T) and meridional wind $(\sim v'T')$ can also be made analogously but is 429 beyond the scope of our study. 430

## 431 **3.3** Extratropical Intrusionintrusion and extreme heavy rainfall events

The next question is, how are the extratropical intrusions or the eddy forcing are linked to the extreme high-intensity rainfall events? Based on We can have two possibilities based on the above two sections (3.1 and 3.2), we can have two possibilities.). First, a convergence of eddy flux can be interpreted as a positive-, i.e., cyclonic curl directly introduced in the mean flow (eq.12 and 13 of Williams et al., 2007). In a monsoon environment, this can invigorate the existing monsoonal convergence directly over a region where the curl is created

438	(such as when a monsoon trough exists over the foothills of the Himalayas). The second
439	possibility is through an ageostrophic secondary circulation mechanism. The generation of
440	cyclonic (anticyclonic) vorticity atin the lower tropospheric regiontroposphere (transverse
441	circulation dynamics) is associated with the formation of surface lows and indirect circulation
442	(e.g., Fig.2 of Hoskins 1982). Such circulation will lead to the generation of ageostrophic
443	vertical velocity, and depending on local factors (like orography and moisture), extreme
444	precipitation <u>dramatic intensification</u> can occur over the elevated topographies (Uccellini and
445	Johnson, 1979; Vellore et al., 2016a; Hunt et al., 2021)(Uccellini and Johnson, 1979; Vellore
446	et al., 2016a; Hunt et al., 2021), suchleading to extreme precipitation, as over the Himalayan
447	region.a result of a jet streak excitation as shown by Hunt et al., (2021). This ageostrophic
448	motion associated with the indirect circulation can extract the kinetic energy of from the mean
449	flow in the downstream region of <u>the</u> jet exit, where it is super-geostrophic (Lau, 1979).(Lau,
450	1979). Extreme precipitation can result from strong ageostrophic vertical upward motion
451	restoring the static stability. Based on Orlansky and Sheldon(1995) (1995), the
452	quasigeostrophic omega equation can be written as:

453 
$$f_0^2 \frac{\partial}{\partial z} \left( \frac{1}{\rho \partial z} \rho w \right) + N^2 \nabla^2 w = s_g + s' - \dots$$
(4)

454 
$$f_{\underline{0}} \frac{2 \frac{\partial}{\partial z}}{\underline{\rho} \frac{\partial}{\partial z}} \rho w + \underline{N}^2 \nabla^2 w = \underline{s}_{\underline{g}} \pm \underline{s}'_{\underline{q}} - \dots - (4)$$

455 Where, *w* is the vertical velocity,  $\rho$  is the density and  $f_0$ ,  $\underline{f_0}$  is the Coriolis parameter and  $N^2$ , 456  $\underline{N^2}$  is the Brunt-Vaisalla frequency, s' and  $\underline{s'}$  is the source term involving friction, diabatic 24 457 heating, and meridional gradient of Coriolis parameter.  $s_g \underline{s}_g$  is the source term involving 458 geostrophic quantities, (denoted by *g* indicated in the suffix), which <u>SutcliffSutcliffe</u> 459 approximated as: $s_g = 2\frac{1}{2}$ 

where l is a <u>unit</u> coordinate <u>aligned</u> along <u>the</u> thermal wind direction and  $v_g \underline{v}_g$  is the 461 geostrophic wind. From Eq (4), (5), (3), and (1), it is clear that  $\zeta_g$  as well as  $\partial \mathcal{V}_g / \partial z \underline{\zeta_g}$ . 462 <u>as well as  $\partial V_g/\partial z$ </u> can be forced/modulated by eddy forcing with through the generation of 463 vertical velocities.  $\nabla . E \nabla . E$  forces the zonal mean wind, which then changes the vertical shear, 464 subsequently changing the source term  $s_g$ . An extreme  $\underline{s}_g$ . A large increase in vertical shear due to 465 E--vector divergence would then lead to an extreme a proportional increase in vertical velocity, 466 resulting in extreme rainfall. Under the right conditions (e.g., moisture supply), increased 467 vertical velocity can result in an increase in rainfall. The jet will move back to the original 468 position after the Rossby wave has propagated out (or dissipated). The cloud--free conditions 469 restore the horizontal meridional (north-south) gradients of temperature, forcing the jet to move 470 471 back to its normal position. The mechanisms explained in this section is are summarized in Fig.2. The theoretical mechanism as proposed here can be verified through the diagnosis of by 472 diagnosing the direct (mean flow modification e.g., cf eq. 1) and indirect (transverse 473 circulation) mechanism based on the E-vector field will be explored in the next session. 474

#### 475 4.0 The lifecycle of extratropical intrusion during the 2013 Uttarakhand extreme event

In the context of the sequence of events described in sec.3, the following section describes 476 the precipitation and circulation plots to quantify the linear and nonlinear perspectives during 477 this event. To understand the nature of the extratropical intrusion, we plot the first two 478 components of E-Vectors (Eddyeddy asymmetry term  $(\overline{u'^2} - \overline{v'^2})(\overline{u'^2} - \overline{v'^2})$  and the eddy 479 momentum flux transfer term $(\overline{u'v'})$  or the EMF term) area averaged over a domain 480 north of the Uttarakhand (30°-40°N;75°-82°E) region in Fig.3a. The primes are defined with 481 respect to a 3 week the monthly mean 10-30June for June 2013. The plot shows that there is 482 strong eddy asymmetry  $\overline{v'^2} \gg \overline{u'^2} \underbrace{v'^2}_{\nu'^2} \gg \underline{u'^2}_{\nu'^2}$  during 16-17 June 2013 and strong northward 483 transfer of EMF. The mean denotes the monthly mean, and the prime represents the deviation 484 from this mean. Such meridional elongation is significant and consistent with theory in the 485 sense that it indicates southward phase propagation associated with northward momentum 486 487 transfer. This meridional intrusion would be discussed in detail in sec.4.2. (e.g., sec. 3b of (Waterman and Hoskins, 2013) A similar EMF flux but weak eddy asymmetry amplitude are 488 seen during 7-9 June. This implies that both flux transfer and eddy asymmetry could play an 489 essential role during this event. This meridional intrusion will be further discussed in detail in 490 sec.4.2. Next, we will see the synoptic situations during this time. 491

492

493

4.1 Precipitation and circulation during the event: synoptic-scale nature of the event

494	In orderFig.3a shows the strengthening of the eddy asymmetry and EMF index (first
495	and the second components of the E-vector), which helps to understand the low-frequency
496	intrusion over the Indian region (Kalshetti et al., 2021). To understand the synoptic
497	situations we first analyse analyze the precipitation during June 2013. Fig. 3b shows
498	the northward propagation based on 20-80 days Lanczos-filtered TRMM precipitation data
499	during June 2013. It shows clear intraseasonal northward propagation in this band-band.
500	Around 16-17 <sup>th</sup> 13-17 June, the rainfall band has strengthened due to the expected monsoon
501	progression after the onset phase and is also seen in Fig.1. Fig.3a shows the strengthening of
502	the EHF and EMF index, which defines the low-frequency intrusion over the Indian region
503	(Kalshetti et al., 2021). (also refer to Fig.1). Similarly, Fig.4 shows the evolution of the rainfall
504	and the 200hPa wind pattern, which confirms the. The meandering contours confirm the deeper
505	southward movement of the jet stream- <u>staying southward of its climatological mean position</u>
506	during summer (~35N). The temporal evolution plots in Fig.4 show how the blob of
507	precipitation evolves and decays over the Uttarakhand region. It is clear from this plot that the
508	rainfall band spans over larger areas spanning a few thousand kilometressquare kilometers,
509	indicating that the event which caused the extreme high-intensity rainfall is not a purely local
510	extreme event phenomenon and is likely to be correlated with large-scale structures. This large-
511	scale nature is further confirmed from the vorticity plots in Fig.5. The temporal evolution on
512	16 <sup>th</sup> June and 17 <sup>th</sup> June shows cyclonic vorticity advection with strong vorticity bands moving
513	eastward with the <u>additional</u> formation of jet streaks. The plot also shows the (on 16 <sup>th</sup> June, the
	27

514 <u>strongest vorticity shading is in the southward descending branch west of 75°E and on 17<sup>th</sup></u> 515 <u>June it is east of 7°5E in the northward ascending branch). The</u> contours of EMF terms 516 <u>showsshow</u> strong northward (<u>i.e.</u>, positive contours) transport of eddy momentum. Thus, 517 <u>cyclonic vorticity advection could be responsible for creating anomalous eddy momentum</u> 518 <u>fluxes, and hence the Uttarakhand region shows strongermore substantial</u> eddy forcing <del>over</del> 519 during this period <u>as compared to other days in June 20013</u>.

In addition to the extratropical system, as shown in Fig. 3, a low-pressure system iswas 520 also propagating over the Indian region. The low-level circulation and the sea-level pressure 521 pattern are shown in **Fig.6**. The strong low-level systemcirculation is associated with the strong 522 moisture inflow. Also, the surface pressure shows negative anomalies over northwest India and 523 positive anomalies over the eastern part along the Bay of Bengal. Strong northerlies and 524 anomalous high pressure and low-pressure regions are also visible over the Afghanistan 525 region, indicating the existence of upper-level waves-over this region. How is the moisture 526 transport<u>transported</u> in the upper level? We plot the 500-200hPa averaged moisture transport 527  $(uq, vq)_{1}$  and see the strong moisture outflow from the <u>southern side</u>, particularly from the Bay 528 of Bengal low-pressure system towards the Uttarakhand region in Fig.7a-d. The north-westerly 529 moisture transport is strong in the upper level, with the dominant source of moisture isbeing 530 the Bay of Bengal-taking a north-westerly turn during this period. The static stability plots 531 (Fig.7e-h) also showshow a robust north-southstrong meridional gradient with the stability 532 decreasing overnorthwest to southeast over the location of the extreme event location rainfall 533

during 16-17<sup>th</sup> June 2013<u>- (blue shading).</u> Thus, **FigFigs.3-7** confirmsconfirm the presence of both the extratropical system and the tropical system during this period with adequate monsoonal and extratropical influence over the Uttarakhand region—. In the next section, we will elaborate on the southward extension of the extratropical intrusion.

538

#### 4.2 Southward digging of extratropical troughs

To understand the southward digging of the trough, which faces decreased stability over 539 the Indian region, we plot *Fig.8*. The plot has multiple variables, all are-averaged for the days 540 15-17 June 2013-and, which are described below: (a) the shading shows the vertical velocity 541 (w) with positive shading means indicating upward motion, (b) the magenta curve shows the 542 streamlines at 200hPa, (c-) blue contours shows the (positive only) relative vorticity 543 at 200hPa, (d) the black contours shows show zonal winds at 200hPa (positive values are 544 contoured) (e) the green curve shows the 1 PVU contour. The black contours show that the 545 mean jet is shifted over the Indian region and is highly meandering. This meandering indicates 546 trough intrusion, confirmed by the blue vorticity contour in the northern part of Pakistan and 547 Indian region over Kashmir and Leh-Ladakh region. The jet is pushed southward with the 548 southward extending intensified Rossby wave trough. The two Two anticyclones are formed on 549 the southern side of the jetstream, aswhich is evident from by the streamlines (magenta color), 550 and the). The deep cyclonic curvature over the northern flank of India shows that the southward 551 extension is deep with an intense Rossby wave trough. The green-coloured 1PVU curve shows 552

553	that the Rossby wave did not break over the Indian region during this period as the breaking of
554	the Rossby wave induces higher PV (>2PVU) over the Indian region(below 30°N). The strong
555	amplitude of anomalous relative vorticity (blue contours showing positive only values),
556	southward curvature of zonal winds in the figure(~1-2PVU or higher) over the Indian region
557	(below 30°N(Fadnavis and Chattopadhyay, 2017). The strong amplitude of anomalous relative
558	vorticity (blue contours showing positive only values) indicates that the Rossby wave is
559	intense, with no sign of weakening and breaking with jet pushed southward. At the same time,
560	however, there is substantial upward vertical velocity (red shading)) and rainfall (Fig.4) over
561	both the central and the north Indian regionregions. It is interesting to note that the Rossby
562	wave vorticity structure does not directly reach the region where the extreme heavy rainfall
563	event occurred and does not show much tilt in the zonal vertical direction until 600hPa- (plot
564	not shown). Hence, none of the above discussions would indicate the rainfall over the
565	Uttarakhand region <u>occurs</u> as <u>anthe</u> end product of <u>a</u> simple in-situ linear <u>processprocesses</u> such
566	as Rossby wavewaves or monsoon low pressureinduced local convergence (refer Fig.2e.g.,
567	the linear part). This in Fig.2). Can this widespread rainfall-can be explained if we assume that
568	the ageostrophic vertical velocity is generated due to a nonlinear conversion process related to
569	downstream amplification associated with baroclinic conversion (Orlanski and Sheldon,
570	1995)(Orlanski and Sheldon, 1995) or a downstream wave-mean interaction or a combination
571	of both over the Uttarakhand region-? Such nonlinear conversion processes and wave mean

interaction can lead to the conversion of zonal mean to eddy kinetic energy or vice versa andgenerate ageostrophic vertical velocity.

To highlight the strong vertical velocity and the ageostrophic component, the vertical 574 575 velocity over a region including the Uttarakhand is plotted in Fig. 9 as a height-latitude plot. The plot clearly shows extremelarge vertical velocity developing over the Indian region 576 developing as. The development of vertical velocity at this scale can be contributed to by two 577 components, one coming from orographically-forced ascent and the other from wave-induced 578 ascent associated with shears (Teixeira, 2014; Cohen and Boos, 2017). As the upper-level flow 579 is getting predominantly westerly and largely baroclinic (i.e.  $\partial_{..} \partial_{..} \partial_{.} \partial_$ 580 developed from barotropic easterlies (i.e., predominantly easterlies at all levels) before 16<sup>th</sup> 581 June. The, the plot shows the development of strong vertical velocity that is associated with 582 strong vertical shear (zonal wind contours). Such shear development would give rise to 583 unstable waves and vertical velocities (contained in the term sg and s'sg and s' in eq.4). We 584 585 also examine (not shown) the bulk Richardson index (the ratio of Brunt Vaisala frequency to vertical shear). The plot shows that This index and also Fig.7d-f show troposphere above the 586 southern Himalayas is slowly gettingbecomes unstable as the westerly shear zone develops 587 over the slopes of the Himalayas. This generation of ageostrophic vertical velocity also 588 indicates the conversion of mean flow to eddy kinetic energy, as can be seen we demonstrate in 589 590 the next section.

591

#### 4.3 Transient Eddy Mean Interaction

**Fig.10** shows the E-vector divergence of E-vector, i.e.  $\nabla$ .  $(\overrightarrow{E})$  which is  $\nabla$ .  $(\overrightarrow{E})$  (shaded) 592 and the E-vectors. The mean is defined as the 21-day average from 9<sup>th</sup> June to 30<sup>th</sup> June, and 593 the daily transient is defined accordingly as a departure from this mean. Components of E-594 vectors and the divergence are computed after that. The divergence plot is averaged for the 595 period 14-18June201318 June 2013. Divergence is very strong for the period 15-17June17 June 596 2013 (not shown). Also, we superimposed the contour of zonal wind averaged during 15-17th17 597 June to show the location of the jet-stream jetstream during this period. The  $\nabla$ . E--vector 598 shadings indicate that over the Uttarakhand region,  $\nabla$ . E is positive, and hence the mean flow 599 gains energy (cf. equequation 1a). The vector plot shows the convergence pattern of the E 600 vector (Ex, Ey). There are two zones of convergence-divergence patterns around 40°N, which 601 are consistent with the locations of blocking ridges (east and west Asian ridges). This transfer 602 of energy is further confirmed from the plot of the  $1/2(u'^2 + v'^2)$ , i.e., local transient eddy 603 kinetic energy (EKE) given by  $1/2(u^2 + v^2)$ ) in Fig.11. The black bar plot from ERA5 604 shows the growth of EKE during that period-and, which reduces afterwardsafterward. The 605 growth of EKE can occur from and the extraction of kinetic energy from the jetstream. The 606 607 growth of EKE increased eddy forcing on the mean can lead to an increase in the upper level zonal flow and vertical shear, as shown in Fig.8. This can lead to the development of indirect 608 ageostrophic circulationvertical velocity over the Uttarakhandforcing region in the 609 presence(eq.5), and with the appropriate feedback of moisture supply from monsoon 610 32

611 intraseasonal oscillations (**Fig.7**)), the condition may explain then be more favorable for the 612 vigorous development of extreme events heavy rainfall episodes over this region. This analysis 613 demonstrates that the <u>wave-mean to eddy conversionflow interaction</u> in the presence of 614 moisture can cause the extreme create favorable conditions for high-intensity precipitation 615 event events, depending on the potential of wave-mean interaction during the intrusion event 616 forcing the zonal wind at the upper level thereby developing shear source in the omega 617 equation(eqn.5) resulting in localized updrafts and extreme rainfall events.

618 5.0 Operational Model Forecast

It is clear from Fig.1 that the operational models (IITM-ERPAS, UKMO, and the ECMWF 619 have failed to capture the event. Among the three models, ECMWF performs better the best. It 620 is natural to evaluate the model performance in the light of the above discussion and see if the 621 relative success or failure is linked to the inability to capture the intrusion or not. The 622 verification of rainfall is plotted in Fig. 12. The plot shows that all thethree models 623 underestimated the rain over the Uttarakhand region. IITM-ERPAS shows low-intensity 624 625 rainfall over the whole of north India; the UKMO rainfall shows a high-intensity rainfall blob over the Pakistan region but missinghas very little over the Uttarakhand region. The ECMWF 626 model, however, shows some success in capturing the rainfall, but it started one day later than 627 the actual event. Thus, rainfall shows a spatial and temporal shift, which could indicate either 628 improper eddy forcing onto the mean flow resulting in improper vertical velocity or improper 629

630	moisture transport or a combination of both. The spatial pattern of moisture transport averaged
631	between 500-200hPa is shown in Fig. 13, which shows that the models fail to forecast the low
632	pressure-induced moisture transport. Improper moisture transport means the moisture is not
633	available for conversion to rainfall. During 15-16June2013 16 June 2013, IITM-ERPAS
634	showshas the cyclonic circulation shifted towards the Bay of Bengal, while for UKMO-shows,
635	it ishas shifted more towards the Arabian Sea, both missing the Indian landmass-, which could
636	cause rainfall over the Pakistan region. The ECMWF model shows two anticlockwise cyclonic
637	circulations over the Arabian Sea and the Bay of Bengal during 15-16 June 2013. Finally, we
638	show the $\nabla$ . $(\overrightarrow{E})$ $\nabla$ . $(\overrightarrow{E})$ in <b>Fig.14</b> to show determine how well the models have forecasted the
639	eddy meansmean interaction and compareby comparing it with the reanalysis data. The top left
640	panel is shown for reanalysis which is like, as in Fig. 10. The reanalyses reanalysis plots show
641	that $\nabla . (\vec{E}) \nabla . (\vec{E})$ is positive (red shading) over the Uttarakhand and Kashmir regionregions,
642	implying acceleration of the zonal mean wind. The hindcast from the models showshows
643	weaker E-vector divergence (red shades) over the Uttarakhand region. Hence, the plot
644	indicates that the eddy E-vector convergence and divergence are not appropriate correctly
645	simulated in the model forecasts over the Uttarakhand region, which is also indicated by the
646	yellow vectors (Ex, Ey) in yellow shadings.). Also, the isotachs are much less dense with
647	lessera less wavy and intruding pattern over Kashmir, Uttarakhand, and Punjab region
648	(between 25°-35°N and 70°-80°E) in the model than the observation reanalysis over the Indian
649	region, implying that the mean flow is also not propercorrectly simulated in the model. Thus,

650 the location of eddy-transient forcing is also not appropriate correctly simulated in the model. Also, the vector shows weaker amplitude of E-vector divergence in the extratropical region 651 north of 30N, implying inappropriate (lower amplitude) eddy transport in the northern 652 extratropics. Due to inappropriate eddy forcing, the zonal wind height longitude profile also 653 does not show the development of appropriate zonal wind acceleration at the upper level (not 654 shown). Inappropriate zonal wind acceleration at the upper-level weekendweakened the 655 vertical shear, thus weakening the <u>SuteliffSutcliffe</u> term in <u>eqnequation</u> (5), indicating weaker 656 ageostrophic vertical velocity, resulting in reduced precipitation amplitude. The result-thus 657 suggest, therefore, suggests that the model did not capture the Uttarakhand event because of 658 inappropriate moisture transport from the monsoon flow and forecast of the wrong location of 659 660 eddy transport leading to thea spatial phase shift of eddy flux divergence and (north and south of Uttarakhand region), resulting in improper eddy forcing over the Uttarakhand region. This 661 is also confirmed from the eddy kinetic energy (EKE) plot in **Fig.11**, which shows EKE plots 662 for the model forecasts. From **figFig.11**, it may be noted seen that the ECMWF, though, model 663 664 captured this time sequence of EKE evolution more effectively compared tothan either the IITM-ERPAS and or the UKMO model models. Thus, the analysis suggests that inappropriate 665 eddy forcing and improper representation of EKE in the model could be a reason for erroneous 666 rainfall forecast in the model. 667

668

669

#### 6.0 Discussion and Conclusion

This case study examines the life cycle of the extratropical intrusion event over 670 Uttarakhand (India) during June 2013 to understand the role of anisotropic shape, meridional 671 propagation of extratropical eddies, and their feedback onto the monsoon mean flow during the 672 2013 Uttarakhand extreme event. Previous studies of extreme events such as those over the 673 Uttarakhand region during June 2013 neglect the role of eddy-mean interaction in the 674 monsoonal region. The study reveals a potential role of underlying eddy dynamics and the 675 inadequacy in the operational forecast models to capture the eddy dynamics. The E-vector-676 based approach that is adopted here gives a first-hand idea about the eddy forcing mechanism, 677 both in the observation and in the model forecast for the 2013 Uttarakhand extreme heavy 678 rainfall event. 679

On the synoptic scale, extratropical Rossby wave intrusion influenced the 2013-June 680 2013 heavy rainfall event over the Uttarakhand extreme eventregion through the southward 681 extension of troughs and associated southward shift in the subtropical jet pattern. Such 682 synergistic evolutions are documented earlier also (Raman and Rao, 1981; Kalshetti et al., 683 2021)(Raman and Rao, 1981; Kalshetti et al., 2021). At a local scale, along with western 684 Himalayan orography and moisture convergence associated with monsoon lows during this 685 time, the upper-level extratropical trough intrusion imparted a strong eddy forcing, which is 686 evident through the existence of E-vector divergence (Fig. 10) and the conversion of mean 687

688	kinetic energy to eddy kinetic energy (Fig.11). Thus, on 16-17 June, upper-level eddy forcing
689	accelerated the eddy circulation dynamics, and developed <u>an</u> additional shear flow that leads
690	to the extreme high-intensity rainfall event by amplification of amplifying the Sutcliff source
691	term(eqn 5). (equation 5). In some locations, the mesoscale circulation can develop through
692	feedback from the large-scale flow. It can lead to cloudburst-type situations by forming a super-
693	convective system, as discussed in Houze et al. (2000). However, we have not focused on this
694	large scale to mesoscale connection in this study as we emphasized the quantification of eddy
695	forcing. The current analysis, unlike earlier studies, differentiates the eddy and meansmean
696	flow over the monsoonal region during this period. Our analysis only assumes the mean flow
697	and background instability associated with extratropical modes and does not assume tropical
698	depressions as a precondition. During the intrusion event, if there is background instability
699	(which can be provided by eddies of extratropical origin), wave-mean interaction can develop
700	ageostrophic velocity leading to extreme high-intensity rainfall events.
701	The above explanation based on the E-vector approach is then verified in operational
702	model forecasts. Result suggests that ECMWF, ERPAS, and UKMO simulated upper-level
703	eirculation patterns, but E-vector divergent field and their impact on underlying atmospheric
704	states seems to be not in order. E-vector divergence and local eddy kinetic energy is captured
705	in varying degrees of The analysis presented here is based on a single extreme event. Does
706	every extreme event over the north Indian region require extratropical intrusion? It is found
707	that high-intensity rainfall events over northern India (along the Himalayan belt and adjoining 37

708	foothills and plains) can occur in the absence of extratropical eddy forcing. Extreme events can
709	occur purely due to monsoon flow over the Indian region and are common during monsoon
710	season over different areas of India. Is there any apparent difference in the spatial pattern of
711	rainfall when an extratropical intrusion occurs? We have checked for two situations when the
712	standardized anomaly of rainfall over the north Indian region (area averaged over a box of 25°-
713	40°N & 65°-90°E) is more than two standard deviations (i.e., the rainfall is high intensity). In
714	the first situation, we computed rainfall, 200-hPa wind vectors, and eddy flux divergence
715	composites in the presence of strong low frequency southward EMF transfer, and in the second
716	situation, when there is negligible EMF transfer. The low-frequency eddy transfer band is
717	defined by taking a 30-60 day filtered EMF index $(F_L)$ area averaged over a box (30°-45°N
718	& 70°-95°E) which is then standardized. Strong southward transfer cases are selected by
719	identifying days when the standardized anomaly is less than -1. This plot is shown in Fig.15.
720	The figure compares the rainfall, wind vector at 200hPa, the divergence of E vector, and E-
721	vector components for the two extreme rainfall scenarios: with or without eddy forcing. The
722	composite is based on 24 selected cases. The plot shows that there are indeed certain regions
723	in the Himalayas and the foothills of the Himalayas, starting from Uttarakhand and extending
724	towards the east, where high-intensity rainfall occurs when there is southward momentum
725	transfer (Fig.15a). High-intensity precipitation also occurs when there is negligible eddy
726	transfer (Fig.15b). The rainfall bias plot (Fig.15c) in the last row shows a positive rainfall
727	anomaly in Uttarakhand. Although there are regional variations, many locations over the
1	38

728	foothills of the Himalayas in the eastern side of Uttarakhand show positive rainfall anomalies.
729	The wind vector plot shows that the cyclone-anticyclone pattern is much closer and stronger
730	over the north Indian region in the strong low-frequency eddy transfer case. The E-vector and
731	its divergence $\nabla$ . $(\overrightarrow{E})$ are shown for the two corresponding scenarios in <b>Fig.15d</b> and <b>Fig.15e</b> .
732	Strong positive values of $\nabla$ . $(\overrightarrow{E})$ indicate substantial divergence in the low-frequency band and
733	strong forcing on the zonal wind over the Indian region (in the Kashmir region). The bias plot
734	for divergence in Fig.15(f) also shows that these extreme events associated with momentum
735	flux transfer require strong E-vector divergence above the Kashmir region. Although we have
736	highlighted the role of EMF and its southward transfer in the low-frequency mode only, all E-
737	vector components can generate extreme events. A detailed analysis for all the bands is required
738	to understand the full implications and it will be reported in a later study. Our initial results
739	based on a case study and a composite analysis in the low-frequency band suggests the
740	usefulness of the E-vector approach in understanding extratropical intrusion event.
741	The above explanation, based on the E-vector approach, was then verified in
742	operational model forecasts. Results suggest that ECMWF, ERPAS, and UKMO operational
743	models simulated upper-level circulation patterns but that the E-vector divergent field and their
744	impact on underlying atmospheric states were not correctly simulated. E-vector divergence and
745	local eddy kinetic energy are captured with varying accuracy, suggesting improper eddy
746	forcing in S2S forecast models. Also, the moisture transport in the upper level is also not
747	captured adequately ( <b>Fig.13</b> ). Thus, inappropriate eddy forcing leads to inappropriate improper 39

748 ageostrophic and mean flow adjustment in the forecast, and inappropriate moisture transport weakens the support of moisture support, leading to a lack inof skill in the model forecast. 749 Although previous analyses provided a description of the event as a Rossby wave intrusion 750 process, the exact role of Rossby wave dynamics and the eddy-mean interaction was not very 751 clear. Our analysis uncovers a series of dynamic steps to understand the reason on-why the 752 753 rainfall is underestimated in operational models during the Uttarakhand extreme heavy rainfall event. Our analysisstudy also provides a diagnostic basis for evaluating the model skill using 754 the E-vector approach and can be used for model skill evaluation. 755

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768 community. IITM CFS model extended forecast runs are available in the IITM data server and 769 would be made available on request (https://www.tropmet.res.in/monsoon/). KHKMRH is funded through Weather and Climate Science for Service Partnership (WCSSP) India, a collaborative 770 771 initiative between the Met Office, supported by the UK Government's Newton Fund, and the Indian Ministry of Earth Sciences (MoES). Authors The authors would like to acknowledge Dr. Gill 772 ,r l. Martin, UKMO, Program Manager Indo-UK WCSSP India project (UK side)), for providing 773 detailed comments and suggestions in this version of the manuscript. 774

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# 978 **Table 1** The important configuration of three S2S operational models used in the

- 979 study. The range stands for forecast lead time in days, resolution represented in
- 980 degrees of latitude versus longitude; letter "L" stands for vertical resolution, ens
- 981 size for ensemble size of the run, frequency of forecast suggests operational run
- 982 scheduled, forecast are past run using output from the actual forecast for
- 983 calibration purpose, ocean, and ice coupled through their respect to the

		Model	Range	Resolution	Frequency	Ens Size		
		Widder					Frequency	-Period
		ERPAS: Extended Range Prediction For Application to Society	<del>Days 0-33</del>	<del>T382 &amp; T126, L6</del> 4	Weekly	<del>16</del>	<del>On the fly</del>	- <u>2003-201</u>
		UKMO: United Kingdom Meteorological Office	<del>Days 0–60</del>	<del>~0.5° × 0.8°, L85</del>	Weekly	4	<del>On the fly</del>	- <del>1996-20(</del>
		EMWF: European Centre for Medium-range Weather Forecasts	<del>Days 0–46</del>	0.25° × 0.25° days 0-10 0.5° × 0.5° After day 10, L85	<del>Twice in week</del>	<del>51</del>	<del>On the fly</del>	Last 20 ye
	984	<del>dynamical core.</del>		9				
	986							

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#### **Graphical Abstract**

Eddy transport, Wave-mean flow interaction, and Eddy forcing during the 2013 Uttarakhand Extreme Event in the Reanalysis and S2S Retrospective Forecast Data

Mahesh Kalshetti<sup>1,3</sup>, Rajib Chattopadhyay<sup>1,2</sup>, Kieran M R Hunt<sup>4</sup>, R Phani<sup>1</sup>, Susmitha Joseph<sup>1</sup>, DR Pattanaik<sup>2</sup>, AK Sahai<sup>1</sup>



# Keywords

Extreme events; Extratropical-Tropical teleconnection; Indian monsoon, S2S subseasonal forecasts.

**Caption**: The study proposes a diagnostic framework using E-vectors to analyze and understand the extratropical transient eddy forcing over Indian Region during the monsoon season. The framework can be used both in observation/reanalysis data and operational forecast models. The analysis is based on a case study of the 2013 Uttarakhand (India) extreme rainfall event over the Himalayan region. The study indicates that the transient eddy forcing can force extreme rainfall events in addition to other forcings.
occipation of the second secon