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The Subseasonal Evolution of Indian Rainfall	
Dipole and its Local Impact in Recent Decades	
C. Sarat <sup>1*</sup> , V. Venugopal <sup>2,3</sup> and Sekhar Muddu <sup>1,4</sup>	
<sup>1</sup> *Interdisciplinary Centre for Water Research.	
<sup>2</sup> Centre for Atmospheric and Oceanic Sciences.	
<sup>3</sup> Divecha Centre for Climate Change.	
<sup>4</sup> Department of Civil Engineering,	
Indian Institute of Science, Bangalore 560012, Karnataka, India.	
*Corresponding author(s). E-mail(s): <a href="mailto:sarat@iisc.ac.in">sarat@iisc.ac.in</a> -or- <a href="mailto:csarat48@gmail.com">csarat48@gmail.com</a> ;	
Contributing authors: <a href="mailto:venu@iisc.ac.in">venu@iisc.ac.in</a> -or- <a href="mailto:venu.vuruputur@gmail.com">venu.vuruputur@gmail.com</a> ; <a href="mailto:muddu@iisc.ac.in">muddu@iisc.ac.in</a> ;	

## Abstract

Over the past two decades, the summer monsoon rainfall has shown decreasing trends over central India, with the Indo-Gangetic plains showing the most prominent changes. In particular, Bundelkhand, a subregion in the Gangetic plains, has experienced an increase in the frequency and severity of meteorological droughts. We analyse long-term rainfall data revealing two types of droughts in Bundelkhand: Type-1 droughts that coincide with the large-scale Indian monsoon droughts, and Type-2 droughts, which are localised to Bundelkhand. Our focus in this work is on the spatio-temporal evolution of rainfall during the latter category of droughts. We find that there is a distinct dipole structure in rainfall on a subseasonal scale, characterised by increased rainfall in the western India region and a decreased rainfall in east central India during July and August. Furthermore, we analyse the upper- and lower-level atmospheric circulation changes responsible for the subsidence during Type-2 droughts. The rainfall deficit during July and August appear to be associated with a midlatitude stationary Rossby wave which induces an anomalous anticyclone over western North Pacific which drives easterlies over the south China Sea into east central India, that in turn reduce the regional moisture influx. At the same time, the Rossby wave also induces an anomalous cyclonic circulation over northwest India and Pakistan,

047 increasing moisture convergence from the Arabian Sea, leading to enhanced con-  
048 vection over west India. During September, the high-pressure region over Tibetan  
049 Plateau migrates further southward into central India, increasing (decreasing)  
050 subsidence (convection). Taken together, a combined effect of intraseasonal vari-  
051 ations in the midlatitude jet and the circulation over western North Pacific and  
052 Tibetan Plateau are responsible for the observed rainfall dipole pattern. At the  
053 downstream end, we find that the enhanced soil stress (dryness) owing to reduced  
054 rain may have a role to play in the much-reported depletion of groundwater  
055 storage in northwest India in general.  
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057 **Keywords:** Indian Monsoon, Drought, Rainfall Dipole, Regional climate  
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## 061 1 Introduction

062

063 The Indian monsoon is an annual climate phenomenon that in rainfall from June  
064 to September (JJAS) over India, with the summer season rainfall accounting for 70-  
065 90% of the total annual rainfall over India. It is modulated by interactions between  
066 land, atmosphere and the tropical Indian Ocean at a range of space and time scales  
067 (Webster et al., 1998). Understanding the spatio-temporal variability of Indian sum-  
068 mer monsoon rainfall (ISMR) is essential for forecasting and policy making, as India's  
069 agricultural yield in the Kharif season depends on summer season rainfall (Gadgil and  
070 Gadgil, 2006). Indian monsoon responds to external drivers such as sea surface tem-  
071 perature (SST) patterns, climate change, and low-frequency climate variability. The  
072 monsoon's internal variability is often seen to be responsible for the active and break  
073 spells in rainfall which, in turn, can be modulated by external drivers. There has been  
074 substantial work done over the past several decades to understand the subseasonal  
075 and seasonal variability of monsoon rainfall and its seasonal extremes (i.e., droughts  
076 and floods). In particular, efforts have focussed on identifying slow-varying oceanic  
077 markers that have a stranglehold on monsoon variability, such as the phases of El  
078 Niño Southern Oscillation (ENSO), namely, El Niño and La Niña respectively (Moo-  
079 ley and Parthasarathy, 1983; Goswami, 1998). Furthermore, the positive (negative)  
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phase of the Indian Ocean Dipole (IOD) has been associated with excess (deficit) in seasonal rainfall of ISM. (Ashok et al., 2001; Gadgil et al., 2004). More recently, there is evidence for extratropical influences from the Atlantic Ocean modulating interannual and subseasonal variability of the Indian monsoon. (Rajeevan and Sridhar, 2008; Yadav, 2009; Boers et al., 2019; Borah et al., 2020).	093 094 095 096 097 098 099 100 101 102 103 104 105 106 107 108 109 110 111 112 113 114 115 116 117 118 119 120 121 122 123 124 125 126 127 128 129 130 131 132 133 134 135 136 137 138
On the secular changes front, several studies have documented a decreasing trend in ISMR over central India in the global warming era. While central India and the Indo-Gangetic Plains have experienced an accelerated decrease in the seasonal quantum of rainfall over the past two decades, western India has shown an increase in ISMR, effectively resulting in an apparent dipole pattern in rainfall (Mishra et al., 2012; Jiang and Ting, 2017; Athira et al., 2023). Besides the changes in rainfall manifesting as a dipole, the frequency and severity of extreme daily rainfall events in central and western India have increased significantly during the past century (Goswami et al., 2006; Roxy et al., 2017). Some studies have attributed the ISMR dipole pattern to long-term increase in sea surface temperature (SST) in the western and Equatorial Indian Ocean (Mishra et al., 2012; Roxy et al., 2015; Goswami, 2023). In addition, the negative phases of IOD favour moisture convergence (divergence) over western India (central India), resulting in the ISMR dipole (Behera and Ratnam, 2018). Furthermore, the ISMR dipole pattern seems to be prominent during a La Niña year, preceded by a La Niña in the previous winter, while transitions from winter El Niño to summer La Niña do not appear to be favourable for the occurrence of a dipole in rainfall patterns. These “Rogue La Niñas” may have a role to play in decreased rainfall over central India even when the Pacific is cooler, thereby altering the canonical relationship between ISMR and Pacific (Gadgil et al., 2023; Sharma et al., 2024).	
There is also some evidence that a westward expansion of the west Pacific subtropical high (WPSH) and increased active tropical cyclone activity over the western north Pacific are responsible for decreased ISMR over central India (Darshana et al.,	

139 2020; Chaluvadi et al., 2024). The Tibetan Plateau, located northeast of India, influ-  
140 ences ISMR on an interannual time scale to the extent that monsoon rainfall and the  
141 Tibetan Plateau can be considered a coupled system. At a seasonal level, the heating  
142 over Tibetan Plateau significantly correlates with early and late season rainfall during  
143 June and September when the Somali Jet is relatively weaker compared to the peak  
144 monsoon period of July and August (Rajagopalan and Molnar, 2013). However, there  
145 is some evidence that differential heating patterns over Tibetan Plateau modulate the  
146 low-level monsoon jet (westerlies), thereby influencing regional vapour transport. The  
147 latent heating over Tibetan Plateau strengthens the westerlies, increasing rainfall over  
148 northwest India. At the same time, the vapour transport from the Bay of Bengal into  
149 central India decreases, resulting in the rainfall dipole (Jiang and Ting, 2017). Finally,  
150 the surface pressure patterns over Tibetan Plateau also could be responsible for the  
151 occurrence of ISMR dipole. Increasing surface pressure over Tibetan Plateau inten-  
152 sifies lower-level easterlies in northwestern India while decreasing mid-tropospheric  
153 northwesterlies in East India and the Indo-Gangetic Plains (Singh et al., 2022).  
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155 Most studies have attempted to link the formation of rainfall dipole to long-term  
156 and seasonal changes in SST patterns and seasonal shifts in the tropospheric circula-  
157 tion over western North Pacific. However, very few studies explored the subseasonal  
158 evolution of the rainfall dipole, although they focused on specific years when the  
159 dipole pattern was evident. To that end, in this study, we perform a composite analy-  
160 sis across years to gain insights into the interannual and subseasonal evolution of the  
161 ISMR dipole, as well as perform a diagnostic analysis of the upper level atmospheric  
162 circulation to identify potential local and remote drivers of observed subseasonal vari-  
163 ability leading to the dipole in rainfall patterns. Specifically, as our analysis revealed,  
164 our focus in part is on a smaller region within the Indo-Gangetic Plains, namely, Bun-  
165 delkhand, which appears to have been the most affected region over the past two to  
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three decades. This region also falls in the belt where significant depletion of groundwater has been reported. Thus, our study while investigating the large-scale factors influencing the dipole pattern in rainfall and its subseasonal evolution, also presents an analysis of impact at the downstream end (namely, soil moisture and groundwater) over the Bundelkhand region.

The outline of our work is as follows: Section 2 presents the data - observations and reanalysis - used in our study, along with the methodology. Section 3 presents the main results of our investigation, with much of it focussing on the rainfall patterns during drought years and the plausible pathways that may have led to the subseasonal variability; the last third of this section (Section 3.4) presents how this rainfall variability may have had a bearing on the local hydrology, specifically on changes in soil moisture and groundwater. We conclude by presenting a summary of our main findings, along with a discussion of a few open-ended questions in Section 4.

## 2 Data and Methods

## 2.1 Data

We use a suite of datasets - both observational and reanalysis - for our study: observed rainfall, soil moisture and groundwater, along with dynamical variables from reanalysis describing the state of the upper atmosphere for the diagnostic part of our analysis. The focus is on June to September monsoon months, referred to henceforth as JJAS.

For rainfall, the daily gridded rainfall at 1° resolution for 1901-2023 from the Indian Meteorological Department (IMD) ([Rajeevan et al., 2006](#)), along with the All India area-weighted mean Summer Monsoon Rainfall (AISMR), developed by the Indian Institute of Tropical Meteorology (IITM; available at <https://www.tropmet.res.in>), are used. The latter dataset is based on a homogeneous rainfall record from 306 rain gauges across India. We use this dataset primarily to identify all-India droughts, as it

231 is considered to be a reliable index of AISMR variability. (Parthasarathy and Mooley,  
232 1978; Parthasarathy et al., 1994),

234 On the diagnostic analysis front, we use the fifth generation atmospheric reanalysis  
235 of the global climate from the European Centre for Medium-Range Weather Forecasts  
236 (ECMWF; <https://cds.climate.copernicus.eu>) (ERA5; Soci et al. (2024)). The daily  
237 reanalysis fields for zonal wind ( $u$ ), meridional wind ( $v$ ), vertical pressure velocity ( $\omega$ ),  
238 Geopotential Height (GPH), vorticity and vertically integrated moisture divergence  
239 (VIMD) are obtained at a spatial resolution of  $0.25^\circ$ .  
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241 For analysing the impact of changes in rainfall at the downstream end, namely,  
242 the response of the land surface, the daily profile of soil moisture (in  $kg/m^2$ ) for  
243 2003-2023 was obtained from the Global Land Data Assimilation System (GLDAS)  
244 Catchment Land Surface Model (CLSM) (Li et al., 2020) at a resolution of  $0.25^\circ$ .  
245 The depth to groundwater level (DGWL) data for wells in the Bundelkhand region  
246 was accessed from the Central Groundwater Board (CGWB) on the India - Water  
247 Resources Information System (<https://indiawris.gov.in/wris/>) portal.  
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## 261 2.2 Methodology

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263 Most of our analysis is based on the anomalies of various observables and reanalysis  
264 fields. Specifically, to understand the subseasonal evolution of a given field  $A$ , we define  
265 daily anomalies, for a given day ( $day$ ) in a year ( $yr$ ), at a location  $(x, y)$  during the  
266 summer monsoon months of June through September. These anomalies, denoted by  
267  $A' (x, y, day, yr)$ , are estimated in a conventional way as the deviation from its daily  
268 climatology:

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$$A' (x, y, day, yr) = A (x, y, day, yr) - \frac{1}{N_{yr}} \left( \sum_{yr=1}^{N_{yr}} A (x, y, day, yr) \right) \quad (1)$$

From these daily anomalies, we construct  $j$ -day anomalies,  $A'_j(x, y, yr)$  as a sum of  
daily anomalies over  $j$  days. 277  
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$$A'_j(x, y, yr) = \sum_{day=1}^j A(x, y, day, yr) \quad (2) \quad 281  
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where  $N_{yr}$  is the available number of years of observations or reanalysis data. The field  
 $A$  in our analysis includes rainfall, wind, geopotential height (GPH) and vorticity. 285  
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Specific to the Indian monsoon context, a drought year is defined by the India  
Meteorological Department, based on the percentage deviation of JJAS rainfall ( $D_i$ )  
from the long-term mean (Eq. 3) or, equivalently, a normalised rainfall deviation ( $Z_i$ ).  
This translates to  $D_i \lesssim -10\%$  ( $Z_i \leq -1$ ) for all-India (Mooley and Parthasarathy,  
1983; Rasmusson and Carpenter, 1983), and  $D_i \leq -17.3\%$  ( $Z_i \leq -1$ ) for Bundelkhand  
(see Table 1). 289  
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$$D_i = \left( \frac{P_i - \bar{P}}{\bar{P}} \right) \times 100\% \quad (3) \quad 300  
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$$Z_i = \left( \frac{P_i - \bar{P}}{\sigma} \right) \quad 302  
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where  $P_i$  is the total seasonal (JJAS) rainfall for a given year  $i$  and  $\bar{P}$ ,  $\sigma$  are its  
long-term mean and interannual standard deviation, respectively. 305  
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### 3 Results

As alluded to in the Introduction, our discussion centres around Bundelkhand region  
in the Indo-Gangetic Plains, where the changes in rainfall appear most prominent.  
This focus is in part motivated by earlier studies that have highlighted a dipole of  
changes in rainfall between central Gangetic plains and west India region (Mishra  
et al., 2012; Athira et al., 2023). A preliminary empirical orthogonal functional (EOF)  
analysis of ISMR clearly illustrates this dipole (Fig. 1). It is worth noting that the  
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323 dipole shows up as the most dominant mode of interannual variability (unlike previous  
324 studies which report this as the 2nd EOF mode). The reason for this is the kind of  
325 normalisation we used. While other studies have normalised rainfall anomalies with  
326 interannual standard deviation, we use the climatology of seasonal standard deviation  
327 for our normalisation, before performing an EOF decomposition (see “Supplementary  
328 Information” for details of normalisation). The rationale for such a choice is that there  
329 is substantial spatial variability in rainfall and to make it homogeneous and more  
330 amenable to an EOF type of analysis, the local seasonal standard deviation is more  
331 apt. In and of itself, this kind of normalisation does not alter the larger, essential  
332 message, i.e., that there is a dipole in rainfall changes: it only serves to amplify the  
333 strength of the dipole.  
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341 A stand-out feature from this preliminary analysis is an apparent regime shift in  
342 the principal component of the first EOF mode (PC-1) shown in Fig. 1b. Prior to  
343 the turn of this century, PC-1 is predominantly negative. However, post-2000, the  
344 PC-1 time series transitions to a positive phase with a marked increase in amplitude  
345 and variance. This shift strongly suggests that the dipole pattern of rainfall (taken in  
346 conjunction with the opposite signs of EOF loading in Fig. 1a), has become a more  
347 frequent and intense mode of interannual variability in recent decades. This finding  
348 provides further justification to focus on these localised droughts which have become  
349 more prevalent in the recent two decades.  
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356 From Figs. 1a, b, the region where the reduction in summer monsoon rainfall is  
357 most evident is Bundelkhand, which lies towards northern central India, bordering  
358 the states of Madhya Pradesh and Uttar Pradesh (Fig. S1). Bundelkhand comprises  
359 13 districts in the Indo-Gangetic Plains, contributing to a total area of approximately  
360 73,700 km<sup>2</sup> (top right panel of Fig. S1). The land use in this region is predominantly  
361 agricultural and highly reliant on rainfall. The average annual rainfall in this region is  
362 approximately 960 mm, with JJAS rainfall contributing nearly 90%. Bundelkhand is  
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infamous for its severe water scarcity due to frequent droughts and ineffective water management strategies. The water scarcity in this region has resulted in poor socioeconomic conditions and lowered agricultural yields (Sharma, 2023; Patidar et al., 2024). To better understand the zonal asymmetry in rainfall, manifesting as a dipole, we focus on this region ( $22\text{-}27^{\circ}\text{N}$ ,  $78\text{-}83^{\circ}\text{E}$ ; dashed rectangle in Fig. 1a) for in-depth analysis.

### 3.1 Drought Categories

To identify droughts, we compare the interannual anomalies of the all-India summer monsoon rainfall (see Section 2.1) with those of rainfall over Bundelkhand, our region of interest (Fig. 2). Between 1901 and 2023, India experienced 24 droughts (Fig. 2a; years when the departure from mean is less than -10%). During this same period Bundelkhand experienced 20 droughts (Fig. 2b). The clustering of droughts in Bundelkhand over the past two decades is evident, with the region experiencing severe rainfall deficit, on an average, once every two years from 2002 to 2020. In both panels of Fig. 2, the “blue” bars represent those years when *only Bundelkhand* experienced severe deficit, while the “red” bars represent Bundelkhand droughts that coincide with large-scale Indian monsoon droughts. In order to understand the features of these “isolated” and localised droughts in Bundelkhand, we consider the following two categories:

- Type-1 Drought: A drought year over India coincides with one in Bundelkhand.
- Type-2 Drought: A drought year exclusive to Bundelkhand.

Bundelkhand experienced 12 Type-1 (“red” bars in Fig. 2) and 8 Type-2 (“blue” bars in Fig. 2) droughts during 1901-2023 (see also Table 2). Type-1 droughts are in fact a subset of the large-scale ISMR droughts that have been studied extensively, with a majority of these years linked to El Niño (Krishnamurthy and Goswami, 2000; Timmermann et al., 2018; Kiran Kumar and Singh, 2021; Ratna et al., 2024). Our focus in this work is on Type-2 drought events, which are localised to Bundelkhand.

415 Curiously enough, Type-2 droughts appear to coincide with weak to moderate La Niña  
416 like conditions in equatorial Pacific (Fig. S2).  
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### 421 **3.2 Dipole Pattern of Rainfall**

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#### *424 Seasonal Footprint*

425 Fig. 3 shows the spatial distribution of seasonal (JJAS) rainfall climatology and com-  
426 posites of the corresponding anomalies for both drought types, respectively, during  
427 the period 1901-2023. As expected, Type-1 droughts (Fig. 3b) are characterized by  
428 large-scale rainfall deficits that engulf a significant portion of the country. Specifically,  
429 Type-1 droughts result in an average seasonal rainfall deficit of more than 150 mm,  
430 primarily concentrated over central India and the Indian western coast. Furthermore,  
431 other parts of the country (e.g., peninsular India), experience an average seasonal rain-  
432 fall deficit of approximately 100 mm, whereas northeast India experiences normal to  
433 above-normal rainfall. This northeast / rest of India contrast in excess/deficit rainfall  
434 is a well-known feature during El Niño droughts, particularly during the past century  
435 (Saikranthi et al., 2018; Kiran Kumar and Singh, 2021). In contrast, Type-2 droughts  
436 (Fig. 3c) are characterized by rainfall deficit that is confined primarily to Bundelk-  
437 hand and a few parts of northeast India. During a Type-2 drought, while central  
438 India receives near-normal rainfall, the western and southern parts of India experience  
439 above-normal rainfall (broadly consistent with the EOF analysis presented in Fig. 1).  
440 This dipole pattern of decreased JJAS rainfall over Bundelkhand and concurrently  
441 increased rainfall over western and southern India is the primary distinguishing factor  
442 between Type-1 and Type-2 droughts.

443 An additional way to characterise this distinction between the two categories of  
444 droughts is the accumulation of rainfall deficit in Bundelkhand, along with the west  
445 India sector (WI: 17-25°N, 69-78°E; see larger dashed rectangles in Figs. 3a, c). The  
446 average build up of rainfall deficit in these two regions (solid curves) along with their

intra-drought spread (shading) are shown in Fig. 4. It is quite clear, and not surprising, 461  
that when the droughts are large-scale in nature (Type-1) rainfall deficit is seen nearly 462  
everywhere (declining solid red and black curves in the left panel of Fig. 4). Contrast 463  
this with the case of localised droughts, shown in the right panel of Fig. 4, one can 464  
readily see the temporal behaviour of the aforementioned dipole structure: a gradual 465  
buildup of rainfall deficit in Bundelkhand (solid blue curve), while the west India region 466  
(solid black curve) shows above normal rainfall with a particularly rapid increase over 467  
a period of 45 days, spanning from mid-July to end-August. 468

The intra-category heterogeneity seen in Fig. 4b is also evident in the spatial 469  
distribution of rainfall anomalies for each of the Type-2 drought years (Fig. S3). 470  
These spatial patterns point to the possibility that the east-west rainfall dipole may 471  
be a recent phenomenon, i.e., more apparent in the Type-2 droughts of the past two 472  
decades compared to those that occurred during the early 20th century. **Indeed, this** 473  
**observation is consistent with the post-2000 regime shift identified in the principal** 474  
**component of EOF Mode 1 (Fig. 1b).** In fact, a comparison of the contrast between 475  
the spatial patterns of rainfall during pre- and post-2000 Type-2 droughts (Fig. 476  
S3a-c *vs.* Fig. S3d-h) also appears to highlight this transition, although the sample 477  
size for either pre- and post-2000 droughts is arguably not large. This suggests that 478  
perhaps the dipole structure may have become prominent only in recent decades. For 479  
example, the rainfall deficit region was mostly confined to Bundelkhand during 2006 480  
and 2007. In these two years, Bundelkhand experienced an average rainfall deficit of 481  
greater than 250 mm, while the surrounding region experienced excess rainfall of a 482  
similar magnitude. In 2010, the dipole shows a marginal shift to the east, resulting 483  
in Bundelkhand, east and northeast India experiencing a rainfall deficit. Also, in the 484  
same year, the southwestern coast of India, where the climatological rainfall is nearly 485  
2000 mm, experienced a deficit of over 250 mm. Even though the rainfall dipole pat- 486  
tern was slightly subdued in 2017 and 2020, Bundelkhand remains the centrepiece for 487  
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507 rainfall deficit, further solidifying the need to understand this peculiar localisation.

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510 *Subseasonal Variability of the Dipole Pattern*

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512 The discussion, thus far, has centred around seasonal rainfall and its deficit. Given  
513 that much of the monsoon rainfall - be it a burst of rainfall or a period of quiescence -  
514 manifests on shorter timescales, often referred to as “active” and “break” spells (Pai  
515 et al., 2016; Saha et al., 2023), we investigate the subseasonal evolution of rainfall and  
516 its spatial distribution, with particular emphasis on Bundelkhand and its neighbouring  
517 regions. For compactness sake, here, we choose to highlight and discuss the 20-day  
518 accumulation of rainfall anomalies for the two types of droughts (Fig. 5 and Fig.  
519 523). During early June, rainfall is near-normal over most of India (Fig. 5a; see also  
520 the cumulative curves hovering mostly around zero in Fig. 4b). Subsequently, the  
521 first indications of a rainfall dipole emerge between late June and early July, when  
522 Bundelkhand and parts of northeast India experience decreased rainfall (Fig. 5b). This  
523 is, once again, consistent with the separation between an increasing and a decreasing  
524 cumulative rainfall anomalies curve for western India and Bundelkhand, respectively  
525 (black and blue curves between day 20 and 40 in Fig. 4b).

526       During this period, we observe a similar pattern of decreased rainfall over the  
527 western Indian coast for Type-1 (Fig. S4b,c) droughts as well. However, the negative  
528 rainfall anomaly in Type-1 droughts during this period covers a significant part of  
529 central India, unlike Type-2 droughts (compare Fig. S4b with Fig. 5b). During the sec-  
530 ond half of July, the peak monsoon season, the dipole structure of rainfall in Type-2  
531 droughts becomes increasingly evident as we observe a band of decreased rainfall diag-  
532 onally across east-central and northwest India (Fig. 5c). Although a similar pattern of  
533 rainfall deficit appears over central India in Type-1 droughts at this stage (Fig. 5c),  
534 the peninsular Indian regions show contrasting behaviour (increase in Type-2 with  
535 normal to less than normal in Type-1). As the season progresses, the dipole shifts  
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to the east during early August, and the region of decreased rainfall aligns broadly with the monsoon trough's climatological position (Fig. 5d). This eastward shift in the dipole is characterised by decreased rainfall over east-central and northern India and increased rainfall over western and peninsular India. At this stage, western India exhibits increased rainfall with a daily anomaly  $\geq 4$  mm/day. In comparison, Bundelkhand and northeastern India experience decreased rainfall with a daily anomaly of  $\leq -5$  mm/day. Between late August and early September (Fig. 5e), even though the magnitude of rainfall deficit (excess) becomes subdued over Bundelkhand and northeast India (western India), the broader dipole structure is still evident. As the monsoon retreats in the second half of September, the zonal rainfall asymmetry dampens, with east-central India and western India now receiving reduced rainfall (Fig. 5f).

To summarise, Type-2 drought conditions reveal similar trends in rainfall between Bundelkhand and western India until early-to-mid July, followed by a divergence of cumulative curves as the latter region experiences more than normal rainfall, while Bundelkhand rainfall trajectory continues its downward trend, leading to a seasonal shortfall approximating 220 mm. Thus, our characterisation of the drought as having a “subseasonal” nature refers specifically to the evolution of this dipole pattern. While rainfall deficit over Bundelkhand persists through the season, the distinct dipole structure with concurrent excess rainfall over western India is most pronounced during a 6 to 8-week period from mid-July to early September (Fig. 4b). It is this intensification and subsequent waning of the dipole within the JJAS season that constitutes its subseasonal evolution. This timeframe is also consistent with the dominant timescales of monsoon's intraseasonal oscillations, suggesting a potential link between the dipole's evolution and broader subseasonal variability. (It is worth noting that we also analysed the spatial distribution of 10-day accumulation of rainfall anomalies (not shown); apart from the fact that one would expect an increase in variability at smaller timescales, it does not alter the essence of our main finding.).

599 **3.3 Atmospheric Circulation**

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601 In order to understand how the upper atmosphere circulation features we perform a  
602 diagnostic analysis of the ERA5 reanalysis data. Again, given that Type-1 droughts  
603 are more large-scale, and likely driven by El Niño and extensively studied, our focus  
604 here is more on the rainfall deficit localised over Bundelkhand region. It is increas-  
605 ingly being recognised that extreme weather events (seasonal or short duration) across  
606 the Asian summer monsoon domain, such as droughts, floods, and heat waves, are  
607 significantly modulated by the interplay between external forcing (e.g., anomalous  
608 SSTs) and internal atmospheric variability (Deng et al., 2019). This internal vari-  
609 ability includes dynamics like teleconnections linking concurrent extremes in different  
610 regions (Fu et al., 2024) and extratropical Rossby waves influencing summer monsoon  
611 rainfall on intraseasonal timescales. (Borah et al., 2020; Fu et al., 2024).

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613 Here, we focus on late July to early September, the period when the dipole structure  
614 in Type-2 droughts is most prominent; specifically, July 20-31, August 1-15 and August  
615 24 - September 8. For these selected time periods, we find that ERA5 captures the  
616 observed rainfall features quite accurately in terms of the vertically integrated moisture  
617 fluxes (not shown).

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619 We begin with an analysis of geopotential height (GPH), wind and vorticity anom-  
620 alies at select pressure levels, on subseasonal scales during JJAS. The wind anomalies  
621 are shown at all chosen pressure levels; GPH anomalies are shown for 200 and 500 hPa,  
622 and vorticity anomalies shown at the lower levels (700 and 850 hPa). Fig. 6 shows  
623 composites of anomalies of GPH (shading), vorticity (shading) along with wind vec-  
624 tors for the first two time periods (mid July to mid-August) and Fig. 7 shows the  
625 same quantities for the late-August/early-September period. In late July, the circum-  
626 global Rossby wavetrain modulates increased and decreased GPH regions in the upper  
627 troposphere (Fig. 6a; 200 hPa). This results in a region of increased GPH over north-  
628 ern Tibetan Plateau (TP) at the 200 hPa level. However, in the mid-troposphere  
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(Fig. 6b; 500 hPa), the increased GPH over TP expands southwards into a significant portion of east-central India. At this level, we also observe a band of anomalous easterlies stretching from the western North Pacific to eastern India. The western North Pacific anticyclone favours the formation of these easterlies, which flow from the South China Sea zonally into central India, resulting in a mid-tropospheric anticyclone over the region. The vorticity forcing from this mid-tropospheric anticyclone drives a barotropic response, resulting in an anticyclonic cell in the lower troposphere (700, 850 hPa; Figs. 6c, d) over central India. The anomalously strong easterlies at 700 hPa oppose the monsoon southwesterlies flowing into east-central India. Simultaneously, the interaction between these easterlies and the climatological monsoon flow establishes an anomalous cyclonic circulation over northwest India and adjoining regions. This downstream development, where the upper-level wave train appears to force a lower-level cyclonic response, is a key feature distinguishing the dynamics of Type-2 drought events. This circulation enhances the southwesterly flow from the Arabian Sea into western India, promoting moisture convergence and convection there. This results in break-like conditions and rainfall deficit over east-central India, while western India experiences enhanced rainfall, consistent with the vorticity patterns in Fig. 6d. This proposed regional dipole mechanism is notably different from the circulation during Type-1 droughts, which are known to be characterized by a large-scale anomalous anticyclone and widespread subsidence across most of India, consistent with a canonical El Niño response (Fig. S5).

During the first half of August, in Type-2 drought events, the midlatitude Rossby wavetrain at 200 hPa curves southward down to 30°N (Fig. 6e). During this period, the strengthening of the anomalous anticyclone over the Tibetan Plateau (TP) is also evident. This anomalous anticyclonic cell over TP disrupts the flow of monsoon southwesterlies into the Himalayan foothills and east-central India, where the climatological monsoon trough is located at this time of the season. The monsoon jet strengthens

in the mid-troposphere, and the easterlies flowing into central India weaken. However, this strengthened monsoon flow creates significant zonal shear as it interacts with the anomalous easterly flow on the southern flank of the Tibetan High (Fig. 6e,f). This interaction establishes a distinct vorticity dipole in the lower troposphere (Fig. 6g,h), with cyclonic shear enhancing moisture convergence over western India and anticyclonic shear promoting subsidence over east-central India. This circulation-driven vorticity pattern is therefore directly responsible for the pronounced rainfall dipole observed during this period. In contrast, the circulation for Type-1 droughts during August does not show a southward migration of the Rossby wavetrain in the upper troposphere (Fig. S5 e,f). However, during this period, the strengthened southwesterlies in the lower troposphere result in an anomalous band of cyclonic vorticity across India signaling an active phase of the monsoon within the broader drought season (Fig. S5 g,h).

Between late August and early September (Fig. 7), during Type-2 drought events, the upper tropospheric anomalous anticyclone over the Tibetan Plateau gets relatively weaker. Additionally, the anomalous western North Pacific anticyclone at 200 hPa also weakens and expands eastward during this period. However, at the 500 hPa level, the anticyclonic cell and the region of increased geopotential height migrate southward over east-central India, strengthening the anomalous easterlies over the region. In the lower troposphere, this results in a distinct zonal shear which strengthens the anticyclonic vorticity over east-central India inducing moisture divergence. Concurrently, the persistent anomalous cyclonic circulation over northwest India maintains cyclonic vorticity and strengthens moisture-bearing winds from the Arabian Sea. This east-west dipole in vorticity directly drives the zonal rainfall asymmetry contributing to the observed rainfall dipole pattern until mid-September. This subseasonal migration and evolution of regional circulation cells in Type-2 events differs from the persistent,

large-scale subsidence that characterizes Type-1 droughts from late August to early September (see Fig. S6). 737  
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To further understand the vertical structure of the atmosphere during Type-2 droughts, we analyse changes in the local Walker (averaged over the Indian latitudinal zone of 17-27°N) and Hadley (averaged over the Indian longitudinal zone of 69-83°E) circulations for those periods when the rainfall dipole is most prominent (Fig. 8). 740  
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This vertical circulation is evidently different from the large-scale, widespread subsidence observed across the entire Indian subcontinent during Type-1 droughts (see Fig. S7 for comparison). This circulation is also concurrent with a zone of anomalous subsidence over the equator that extends further to 10°S. When the zonal asymmetry in rainfall dipole starts weakening in early September, we observe an anomalous shallow descending motion to the west of 75°E, indicating decreased convection over western India. During this period, the deep subsidence in the 75-100°E longitude band indicates decreased convection over east-central India. Around the same time, the local Walker circulation appears to shift southward, evident in the strong ascent from the equator to 10°N (Fig. 8c, d), indicating increased convection to the south of India. As a result, peninsular India now experiences near-normal rainfall in early September instead of increased rainfall (as was the case during August). However, during this early September period, the 20-30°N latitude band still experiences anomalous descending motion, consistent with

783 decreased rainfall over north India (including Bundelkhand). Overall, this sub-seasonal  
784 evolution aligns well with the seasonally averaged (JJAS) vertical velocity anomaly  
785 patterns observed during Type-2 droughts (Fig. S8b, d). The composites clearly show  
786 the characteristic mid-tropospheric dipole over India, with strong anomalous subsi-  
787 dence over east-central India and ascent over western and peninsular India (Fig. S8  
788 b). Furthermore, weak anomalous ascent is seen over the Maritime Continent and  
789 Western Pacific, indicating a distinct large-scale circulation pattern.  
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### 797 **3.4 Effects on local hydrology**

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799 Thus far, we have investigated the upper atmospheric circulation features, and plau-  
800 sible pathways leading to localised droughts in the Bundelkhand region, which appear  
801 distinctly different from conventional large-scale Indian droughts in terms of the  
802 space-time structure of evolution of rain. Equally, if not more, importantly, such meteo-  
803 rological droughts can potentially have a significant impact on the region's agricultural  
804 output, threatening food security and overall socio-economic stability. In particular,  
805 the response of soil moisture to rainfall is highly non-linear, and the propagation of a  
806 meteorological drought into what one might deem a "soil moisture drought" depends  
807 on antecedent soil moisture conditions ([Zhu et al., 2021](#)). Understanding the soil mois-  
808 ture variability during June–October is therefore essential, as it is the Kharif crop  
809 season for agriculture in India. To this end, we quantify the footprint of these two  
810 categories of droughts on soil moisture, which typically exhibits a temporally lagged  
811 response to meteorological droughts due to the inherent buffering capacity of soils.  
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814 Fig. 9 shows the composites of normalised monthly soil moisture anomalies during  
815 Type-1 (top row) and Type-2 (bottom row) droughts for the period July to October,  
816 to account for the lagged response of the land surface to rain, at least as seen in  
817 soil moisture. The distinction between the two types of droughts on the downstream  
818 end, namely, in terms of soil moisture patterns, is even more striking than in the  
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rain patterns discussed earlier. In particular, during Type-1 droughts, we reported  
that most of central India receives significantly less than normal rain during JJAS;  
correspondingly, the soil moisture deficit is spread across a large part of the Indo-  
Gangetic Plains (Figs. 9a-d). Additionally, we also observe increased soil stress (i.e.,  
decreased soil moisture) in the west India region, and peninsular and southern India  
regions, as they too receive less-than-normal rainfall (see Fig. 3b or Fig. S4). The soil  
stress persists well into the post-monsoon period (negative soil moisture anomalies  
in October in Fig. 9d) in these climatologically dry western and peninsular Indian  
regions, mainly due to the cumulative effect of decreased rainfall during JJAS (see, for  
instance, the end of the black curve corresponding to west India in Fig. 4a). However,  
the southern tip of India shows elevated soil moisture in response to rainfall during  
the onset of the northeast monsoon in October when the east coast and peninsular  
India begin to receive rainfall. Finally, while most of India experiences severe soil stress  
(negative soil moisture anomalies), northeast India sees above normal soil moisture in  
response to increased JJAS rainfall.

In Type-2 droughts, even though central India and the Indo-Gangetic Plains region  
experience marginally decreased soil moisture in July, the deficit is mainly confined  
to Bundelkhand (Figs. 9e-h). Unlike in the Type-1 case, the west and Peninsular  
India regions show increased soil moisture availability in response to increased rain-  
fall during late July. Subsequently, the cumulative effect of rainfall distribution in the  
localised Type-2 drought becomes evident from August as parts of central India and  
Bundelkhand experience decreased soil moisture. In contrast, increased soil moisture  
persists in other regions of the country as they receive near-to-above-normal rainfall.  
Furthermore, the magnitude of soil moisture in Bundelkhand during Type-2 droughts  
is lesser than that during Type-1 droughts. As the season progresses into the month  
of October (i.e., post-monsoon), while Peninsular and west India region show signif-  
icantly increased soil moisture, the deficit remains localised to Bundelkhand. Thus,

875 in addition to rainfall evolution and its spatial extent, the spatial distribution of the  
876 soil moisture anomalies and their magnitude also serve as key differentiators between  
877 the two drought scenarios in a hydrological context, highlighting the need to consider  
878 these drought categories in a more holistic framework, for improved water management  
881 strategies.

883 Meteorological droughts also influence regional groundwater (GW) dynamics. Like  
884 soil moisture, the response of the regional groundwater to rainfall is also lagged, owing  
885 to the storage characteristics of the underlying aquifer. However, the sustained deple-  
886 tion of groundwater reserves in Bundelkhand is evident in the evolution of the depth  
887 to groundwater levels (Fig. 10) over the past two decades. Between 1995 and 2019,  
888 Bundelkhand experienced five Type-1 droughts (2002, 2004, 2009, 2014 and 2015)  
889 and four Type-2 droughts (2006, 2007, 2010 and 2017). The depletion in ground-  
890 water is particularly prominent during 2005 and 2009 (Fig. 10a) when the region  
891 experienced a meteorological drought nearly every year between 2004 and 2009. Dur-  
892 ing this period, the post-monsoon average depth to groundwater level decreased from  
893 5m before 2005 to almost 10m by the end of 2008, leading to a severe reduction in  
894 the available groundwater storage. Notably, northern Bundelkhand experiences more  
895 drawdown compared to the south; interestingly, this depletion appears to mirror the  
896 spatial heterogeneity in rainfall climatology between the northern and southern parts  
897 (see Fig. S1). This region also exhibits slower recovery post-2009, highlighting the  
898 compounding impact of frequent droughts in an arid situation. Further, a marginal  
899 decline in groundwater levels is observed between 2015 and 2018, coinciding with the  
900 three Type-2 droughts in 2014, 2015, and 2017. This pattern suggests that frequent  
901 meteorological droughts are plausible drivers of subsurface storage depletion in Bun-  
902 delkhand, potentially worsening water availability. However, the role of anthropogenic  
903 adaptations must be considered, particularly in a region like Bundelkhand, which lies  
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in the Indo-Gangetic Plains, where agriculture is highly reliant on GW pumping ([Maurya et al., 2024](#)). Hence, the GW depletion in Bundelkhand is attributable to serial meteorological droughts compounded by concurrent human intervention.

## 4 Summary and Discussion

Over the last two decades, the decrease (increase) of summer monsoon rainfall over central India has manifested as a dipole pattern. Whether this recent decline in rainfall is only climate change related or a combination of secular changes with multidecadal variability still remains an open question. Even though previous studies have analysed droughts during year with a rainfall dipole, their subseasonal evolution has not been explored in detail. In this study, we analyse the June through September (JJAS) rainfall over the past century (1901 to 2023), and show that the rainfall deficit zone during these years of dipole is primarily localised over Bundelkhand, a water-scarce region in east-central India. Based on the spatio-temporal structure of rainfall, we categorised the droughts over Bundelkhand as Type-1 and Type-2. Type-1 droughts - coinciding for the most part with all-India droughts, are characterised by widespread rainfall deficit events encompassing both Bundelkhand and central India. On the other hand, during Type-2 droughts, rainfall deficit (excess) is highly localised to the Bundelkhand (west and peninsular India) region, resulting in what appears to be a prominent dipole pattern in rainfall.

Our analysis further reveals that the subseasonal spatial and temporal evolution of rainfall in these two droughts are distinctly different. In Type-1 droughts, rainfall deficit initially appears across central India and the western coast of India, starting in mid-June and is sustained throughout the JJAS season. In contrast, the rainfall deficit in Type-2 droughts initially develops across northeast India and Bundelkhand around mid-July. The magnitude of rainfall deficit (excess) peaks in Bundelkhand (west and peninsular India) during early August and sustains until early September, signifying

967 the subseasonal nature of the observed dipole pattern. Even though the subseasonal  
968 rainfall evolution differs between the two drought types, the final end-of-season rainfall  
969 deficit in Bundelkhand reaches a magnitude of approximately 300 mm in both cases  
970 (nearly 30-40% of the seasonal accumulation).

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973       The contrasting spatio-temporal rainfall patterns between these droughts  
974 prompted us to analyse the atmospheric circulation. As Type-1 droughts are a subset  
975 of the classical ISMR droughts linked to ENSO, we primarily focussed on the less-  
976 studied Type-2 droughts. We observe that a midlatitude stationary Rossby wavetrain  
977 induces an anomalous anticyclone over western North Pacific between late July and  
978 mid-August. This anomalous anticyclone then drives easterlies over the South China  
979 Sea into east-central India, reducing the regional moisture influx from monsoon north-  
980 westerlies. During August, the midlatitude Rossby wavetrain migrates southwards to  
981 30°N and modulates an anomalous anticyclone over the Tibetan Plateau and along the  
982 Himalayan foothills. As a result, atmospheric subsidence over east-central India leads  
983 to decreased convection and rainfall. During September, the high-pressure region over  
984 Tibetan Plateau migrates further southward into central India, increasing (decreas-  
985 ing) subsidence (convection). The midlatitude westerly Rossby wavetrain also induces  
986 an anomalous cyclonic circulation over northwest India and Pakistan, increasing mois-  
987 ture convergence from the Arabian Sea, leading to increased convection (rainfall) over  
988 west India, thereby contributing to zonal asymmetry (dipole) in rainfall. In summary,  
989 a combined effect of intraseasonal variations in the midlatitude jet and the circulation  
990 over western North Pacific and Tibetan Plateau are responsible for the observed sum-  
991 mer monsoon rainfall dipole pattern. Having said that, further investigation is needed  
992 in this regard as a flood situation in northwest India and Pakistan induces strong dia-  
993 batic forcing, affecting the location of the crucial midlatitude jet ([Fu et al., 2024](#)). To  
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this end, the dynamics of the southward migration of the midlatitude jet on intraseasonal timescales, that induces a Tibetan high during Type-2 droughts, needs further investigation.

Apart from rainfall deficit, meteorological droughts also impact other critical variables within the local hydrological cycle at the downstream end. Specifically, during Type-1 droughts, we find large-scale deficit in soil moisture across central and peninsular India during the peak monsoon months of July and August, persisting into the post-monsoon season. Type-2 droughts, on the other hand, exhibit more localised deficit in soil moisture, which in fact is confined to Bundelkhand (providing yet another distinct difference between the two kinds of droughts). Furthermore, we find that the enhanced soil stress (dryness) owing to reduced rain may have a role to play in the much-reported depletion of groundwater storage in northwest India in general ([Rodell et al., 2009](#); [Bhanja et al., 2016](#)) and Bundelkhand in particular (over and above human intervention), especially in the recent twenty years. Consequently, both Type-1 and Type-2 droughts significantly threaten agricultural yields in Bundelkhand due to sustained soil moisture deficit and diminishing groundwater availability.

In the Indo-Gangetic plains, where the columnar water vapour has comparable contributions from moisture convergence and local evapotranspiration, the regional soil moisture has the potential to modulate local rainfall patterns through positive or negative feedback pathways ([Yeh et al., 1984](#); [Varikoden and Revadekar, 2018](#); [Qing et al., 2023](#)). If positive feedback occurs between rainfall and soil moisture, Type-2 droughts could further decrease (increase) the availability of atmospheric moisture for rainfall in Bundelkhand (western India). Indeed, recent studies document significant increases in precipitation and moisture availability driving widespread greening in parts of Western India, such as the Thar Desert ([Kashyap et al., 2025](#); [Mishra et al., 2025](#)). Interestingly, this greening is sustained not only by enhanced rainfall but also significantly by groundwater extraction, highlighting the complex interplay between

1059 climatic shifts and hydrological management in the region ([Mishra et al., 2025](#)). On  
1060 the other hand, it is plausible that if there is a negative feedback between rainfall  
1061 and soil moisture, increased soil moisture might not result in increased local rainfall  
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1063 over western India. Thus, investigating the soil moisture-atmosphere interactions dur-  
1064 ing Type-2 droughts would be essential to further advance our understanding of the  
1065 physical mechanisms involved in these feedback loops.  
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1067 The findings presented here have important implications for drought management  
1068 in Bundelkhand (and Indo-Gangetic Plains). Any anthropogenic activities, such as  
1069 increased groundwater pumping for agriculture, could only serve to aggravate water  
1070 scarcity concerns. Further depletion of groundwater resources could trigger a positive  
1071 feedback loop: declining soil moisture, coupled with reduced evapotranspiration, can  
1072 decrease atmospheric moisture availability, increasing the risk of localised droughts.  
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1074 Further investigations into local groundwater flow, recharge, and evapotranspiration  
1075 patterns are thus crucial. Finally, drought mitigation strategies for Bundelkhand must  
1076 be informed by the distinct vulnerabilities associated with both Type-1 and Type-2  
1077 droughts.  
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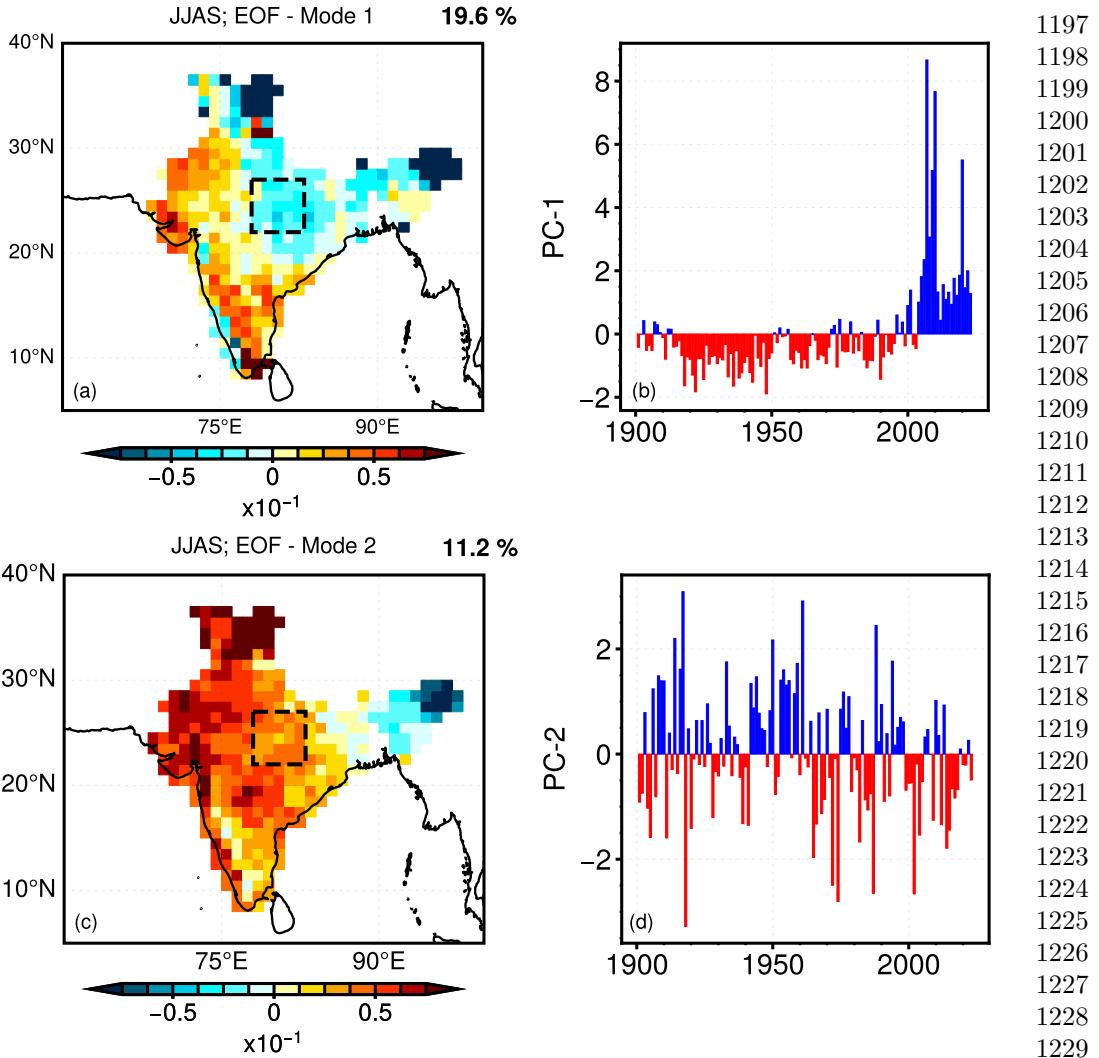
Drought Year	Bundelkhand		
	$D_i$	$Z_i$	
1905	-28.77	-1.67	1105
1907	-24.46	-1.38	1106
1913	-30.11	-1.74	1107
1918	-35.20	-1.42	1108
1928	-32.62	-1.44	1109
1941	-30.12	-1.74	1110
1965	-34.31	-1.99	1111
1966	-23.31	-1.35	1112
1979	-40.06	-2.32	1113
1989	-19.61	-1.13	1114
2002	-19.75	-1.14	1115
2004	-20.03	-1.16	1116
2006	-31.21	-1.81	1117
2007	-29.67	-1.72	1118
2009	-35.62	-2.06	1119
2010	-19.64	-1.14	1120
2014	-33.20	-1.92	1121
2015	-34.30	-1.99	1122
2017	-28.04	-1.62	1123
2020	-17.93	-1.04	1124
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**Table 1:** Percentage seasonal (June - September) rainfall deviation ( $D_i$ ) and normalized JJAS rainfall deviations ( $Z_i$ ) from the long-term mean for all drought years in Bundelkhand. See Eqn. (3).

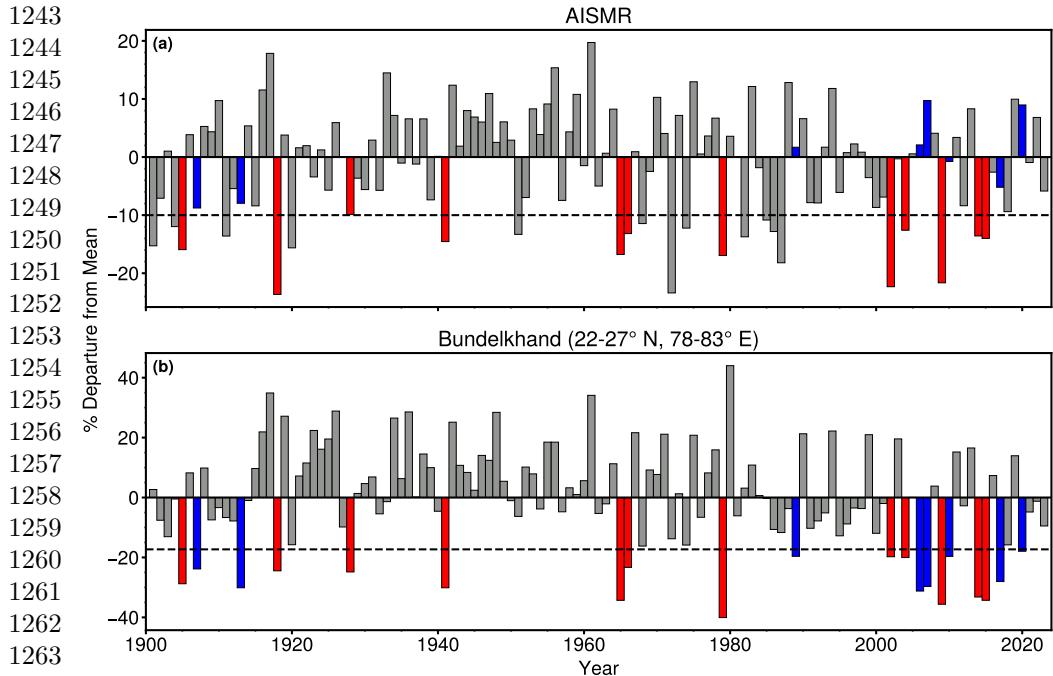
1151	Drought Year	Rainfall Anomaly (%)	
		Bundelkhand	AISMR
1152	1905	-29	-16
1153	1907	-24	-9
1154	1913	-30	-8
1155	1918	-35	-24
1156	1928	-33	-10
1157	1941	-30	-15
1158	1965	-34	-17
1159	1966	-23	-13
1160	1979	-40	-17
1161	1989	-20	2
1162	2002	-20	-22
1163	2004	-20	-13
1164	2006	-31	2
1165	2007	-30	10
1166	2009	-36	-22
1167	2010	-20	-1
1168	2014	-33	-14
1169	2015	-34	-14
1170	2017	-28	-5
1171	2020	-18	9

**Table 2:** Seasonal (June - September) rainfall deviation from long-term mean for Bundelkhand and all India, as a % (Eqn. 3), for each drought year shown in red for Type-1 and blue for Type-2 droughts. These values form the basis for Fig. 2.

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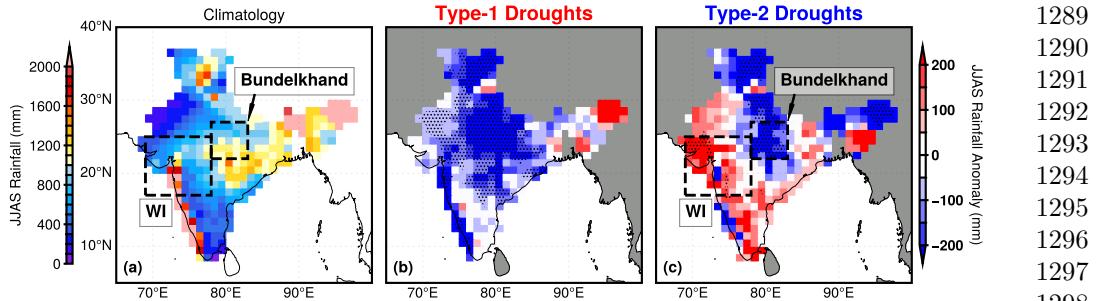


**Fig. 1:** The first two modes of an Empirical Orthogonal Function (EOF) analysis of interannual JJAS rainfall (normalised; see Appendix A) during 1901-2023. (a,c) EOFs and (b, d) corresponding principal components. The region inside the dashed rectangle represents Bundelkhand ( $22\text{-}27^\circ\text{N}$ ,  $78\text{-}83^\circ\text{E}$ ; see also Fig. S1). Based on IMD 1-degree, daily rainfall for 1901-2023.

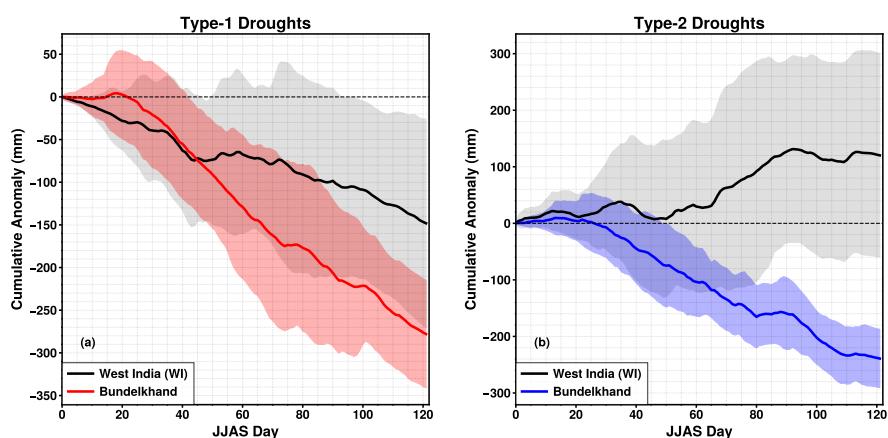


**Fig. 2:** Long-term variation of (a) all-India and (b) Bundelkhand summer monsoon rainfall anomalies, shown as a departure (in %) from long-term mean. As per the India Meteorological Department (IMD), a year is classified as a drought if the seasonal deficit is more than 1 interannual standard deviation. In terms of % departure from the long-term mean, this translates to  $\approx 10\%$  for India as a whole and  $\approx 17\%$  for Bundelkhand, a smaller region (shown as dashed horizontal lines in both panels; see also Tables 1 and 2). The red and blue bars denote Type-1 (common to both regions) and Type-2 (specific to Bundelkhand) droughts. Based on IITM homogeneous monthly rainfall and IMD 1-degree, daily rainfall for 1901-2023.

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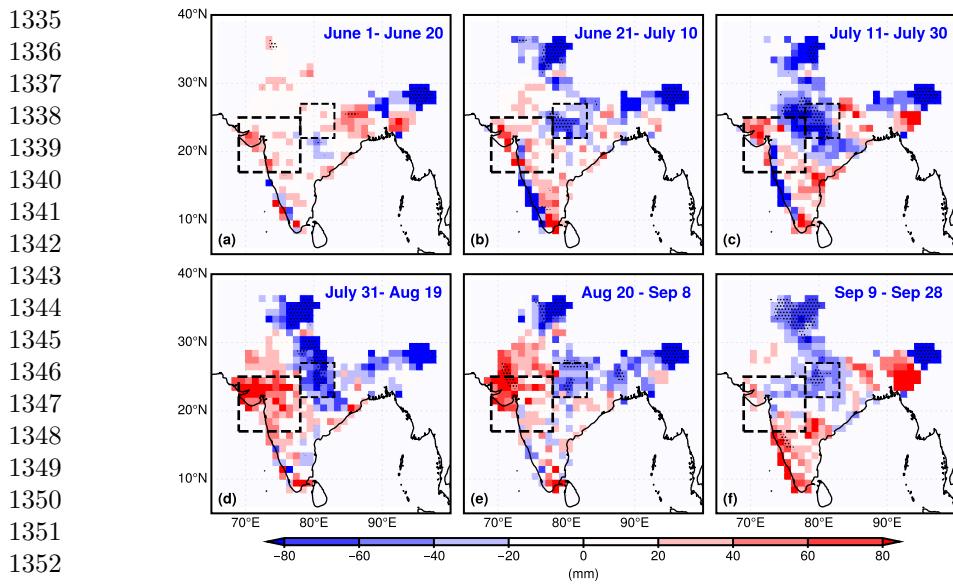


**Fig. 3:** Spatial distribution of June-September rainfall (a) climatology for 1901-2023; and (b, c) anomalies for Type-1 and Type-2 drought composites, respectively. The smaller and larger rectangles denote Bundelkhand and West India (WI) respectively. Stippling indicates regions where the anomalies are statistically significant at the 95% confidence level based on a two-sided Student's t-test. Based on IMD 1-degree, daily rainfall for 1901-2023.



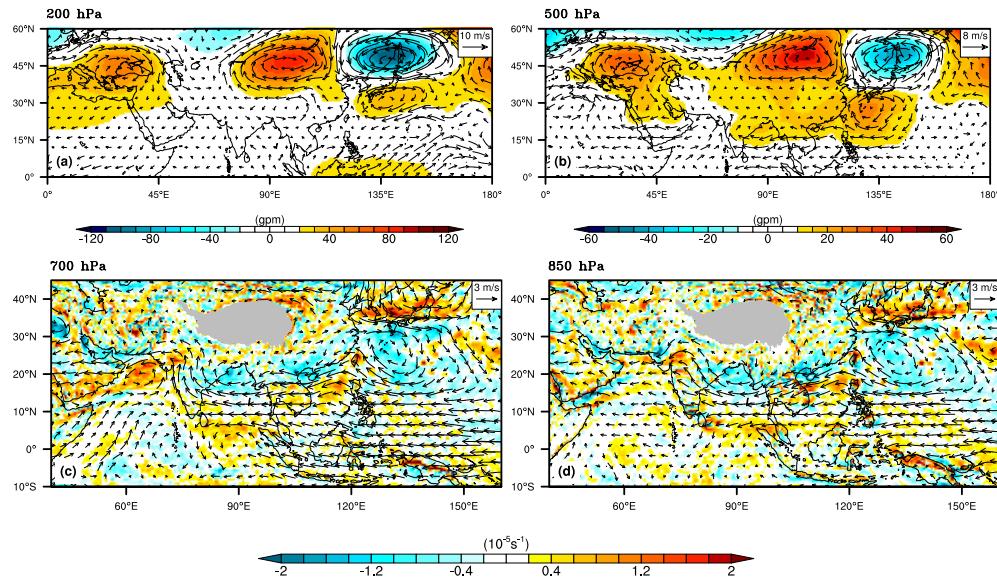
**Fig. 4:** Composite of the cumulative of daily area-averaged rainfall anomalies (mm) for Bundelkhand and western India regions during (a) Type-1 and (b) Type-2 droughts. The solid lines denote the mean cumulative anomaly, and the shading represents  $\pm 1$  standard deviation, estimated from 12 Type-1 and 8 Type-2 drought years. Based on IMD 1-degree, daily rainfall for 1901-2023.

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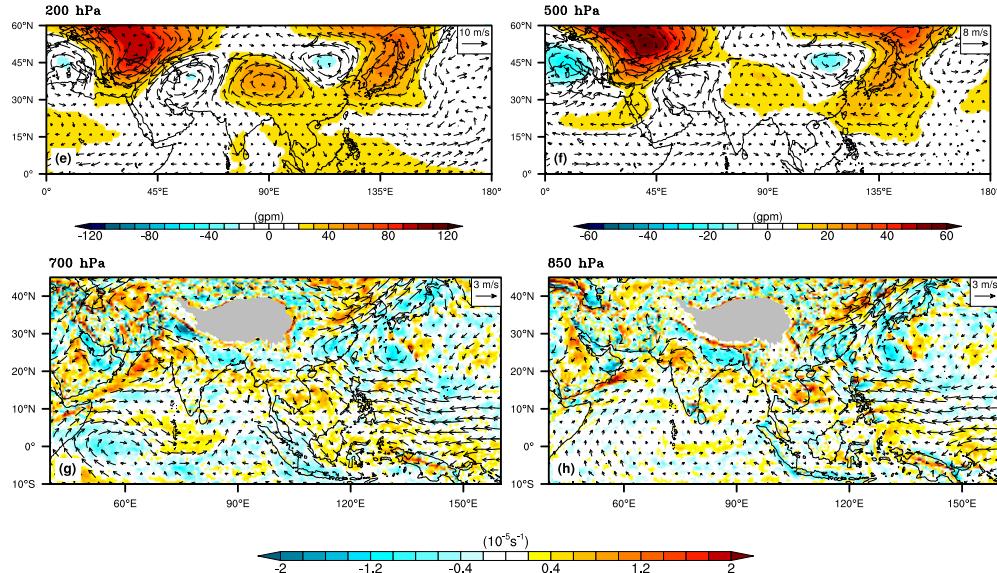


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1353 **Fig. 5:** Spatial distribution of cumulative rainfall anomalies over a 20-day timescale for  
1354 Type-2 droughts. The large and small rectangles denote west India and Bundelkhand  
1355 regions, respectively. Stippling indicates regions where the anomalies are statistically  
1356 significant at the 95% confidence level following a two-sided Student's t-test. Based  
1357 on IMD 1-degree, daily rainfall for 1901-2023.  
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### Type - 2 Droughts : July 20-31

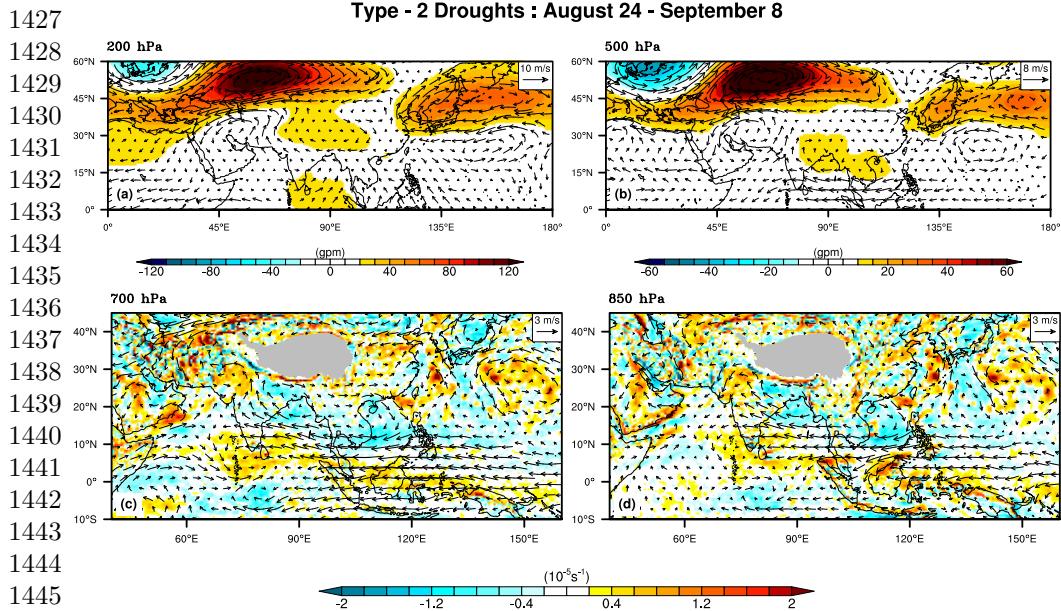


### Type - 2 Droughts : August 1 - August 15



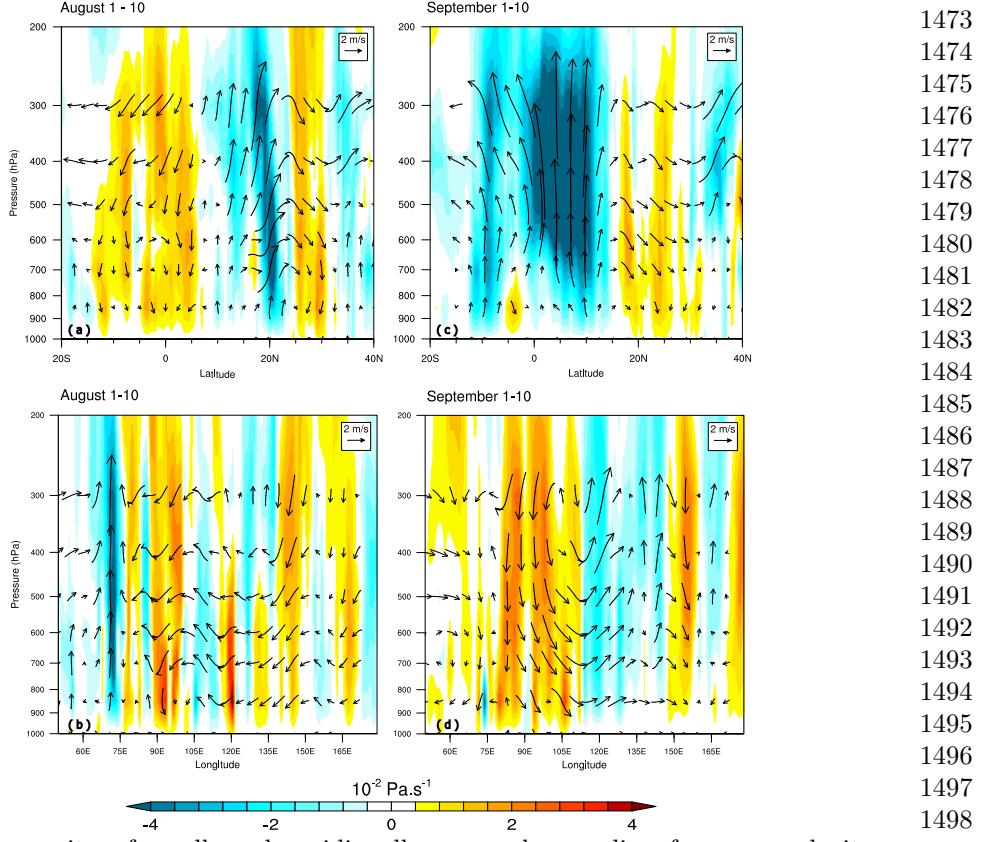
**Fig. 6:** Type-2 drought composites for the periods July 20-31 and August 1-15: (a, b, e, f) Geopotential Height (gpm; shading) and wind anomalies (vectors) at 200 and 500 hPa. (c, d, g, h) wind anomalies (vectors) and relative vorticity ( $10^{-5} s^{-1}$ ; shading) at 700 and 850 hPa. Only the five most recent Type-2 drought years were included in the composites. The shading in grey (in panels c, d, g and h) denotes the Tibetan Plateau. Based on ERA5 0.25-degree, daily reanalysis data for 1970-2023.

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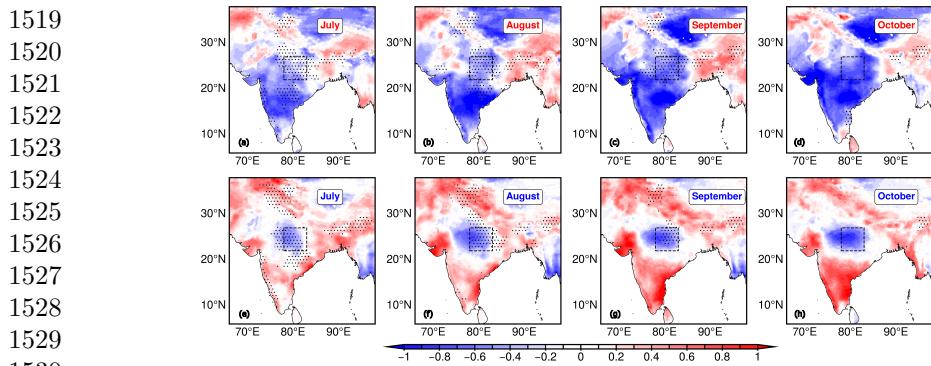


**Fig. 7:** Same as in Fig. 6, but for the period August 24 - September 8.

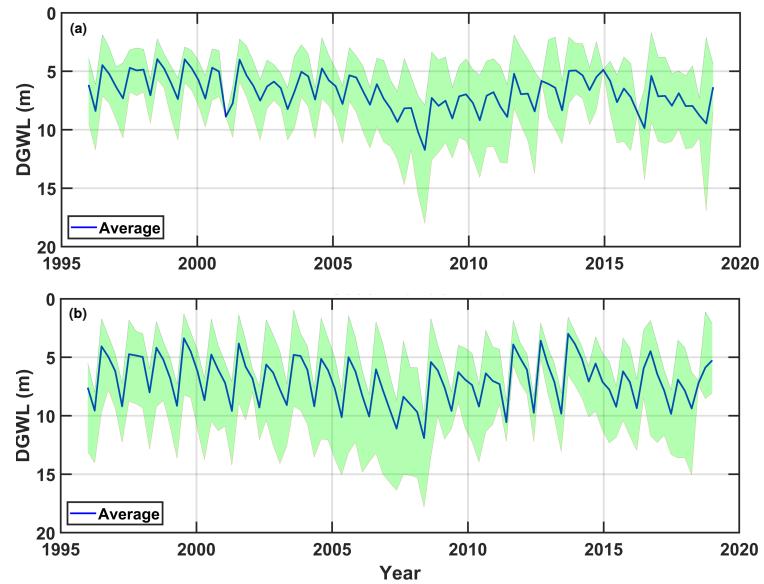
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**Fig. 8:** Composites of zonally and meridionally averaged anomalies of pressure velocity ( $\omega$ ; shading) and wind (vectors) during Type-2 droughts for (left column) August 1–10 and (right column) September 1–10. (a, c) Meridional-vertical and (b, d) Zonal-vertical cross-sections. The zonal and meridional averages are computed over  $69^{\circ}\text{E}$ – $83^{\circ}\text{E}$  and over  $17^{\circ}\text{N}$ – $27^{\circ}\text{N}$ , respectively. Only the five most recent Type-2 drought years were included in the composites. The vertical velocity ( $w$ ) used for the wind vectors has been scaled by  $10^3$  for ease of visualisation. Based on ERA5 0.25-degree, daily reanalysis data for 1970–2023.



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1531 **Fig. 9:** Normalised soil moisture anomalies for (top row; a-d) Type-1 and (bottom row;  
1532 e-h) Type-2 drought composites from July till October. Regions with rainfall deficit  
1533 ( $\leq -1$  mm/day) during the same month are hatched. The normalised anomalies are  
1534 estimated as was done for rainfall. The dashed box denotes Bundelkhand. Based on  
1535 GLDAS 0.25-degree, daily soil moisture and IMD 0.25-degree, daily rainfall data for  
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1560 **Fig. 10:** Time series of depth to groundwater level (DGWL; m) in (a) northern and  
1561 (b) southern Bundelkhand. The blue line represents the mean, and green shading  
1562 represents minimum and maximum DGWL. Based on Central Ground Water Board  
1563 (CGWB) well data for 1995-2019.  
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## 1903 **Declarations**

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1906 **Conflict of Interest.** The authors declare that they have no conflicts of interest.  
1907

1908 **Data Availability.** All data used in this study are publicly available; repository  
1909 links have been provided where applicable. Python and NCL scripts developed for the  
1910 analysis are available from the corresponding author upon reasonable request.  
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1919 1920 Phase III.  
1921

1922 **Author Contributions.** Conceptualization: C. Sarat, V. Venugopal; Methodology:  
1923  
1924 C. Sarat, V. Venugopal; Formal analysis and investigation: C. Sarat; Writing – original  
1925  
1926 draft: C. Sarat; Writing – review and editing: C. Sarat, V. Venugopal, and Sekhar  
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