Precipitation Changes in India

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Key Messages

- There has been a decreasing trend in the all-India annual, as well as summer monsoon mean rainfall during 1951–2015, notably so over areas in the Indo-Gangetic Plains and the Western Ghats. Increasing concentrations of anthropogenic aerosols over the northern hemisphere appear to have played a role in these changes (medium confidence).
- The frequency of localized heavy rain occurrences over India has increased during 1951–2015 (high confidence). Urbanization and other land use, as well as aerosols, likely contribute to these localized heavy rainfall occurrence (medium confidence).
- With continued global warming and expected reductions of aerosol concentrations in the future, climate models project an increase in the annual and summer monsoon mean rainfall, as well as frequency of heavy rain occurrences over most parts of India during the twenty-first century (medium confidence).
- The interannual variability of summer monsoon rainfall is projected to increase through the twenty-first century (high confidence).

3.1 Introduction

Precipitation is an important component of the global water cycle, and the impacts of anthropogenic climate change on precipitation have significant implications on agricultural activities (Porter et al. 2014). Over India, the seasonal monsoon rains during the June-September months, which contribute to more than 75% of the annual rainfall (Fig. 3.1a), are vital for the country's agriculture and economy (Parthasarathy et al. 1988; Gadgil 2007). Abrupt changes in the Indian monsoon precipitation on decadal and centennial time-scales are evident from high-resolution climate proxy records, extending back several thousands of years (Berkelhammer et al. 2012; Sanyal and Sinha 2010). This chapter provides an assessment of historical changes in precipitation, as well as projections of future changes from climate models.

Box 3.1 Processes in the Monsoon System

The annual cycle of monsoon precipitation can be described as a manifestation of the seasonal migration of the Inter-Tropical Convergence Zone (ITCZ) and the associated shift in rain band from southern to northern Indian Ocean (Fig. 3.1a). Summer monsoon rainfall across the Indian Subcontinent shows substantial spatial variability with heaviest rainfall along the Western Ghats and the Himalayan foothills due to orographic features, and over central India due to low-level convergence (Fig. 3.1; upper panel).

Mean Onset and Withdrawal

The onset of summer monsoon over India is characterized by the dramatic rise in mean daily rainfall over Kerala (Ananthkrishnan and Soman 1988; Soman and Kumar 1993). The onset of the Indian summer monsoon (ISM) is a key indicator characterizing the abrupt transition from the dry season to the rainy season and subsequent seasonal march (Koteswaram 1958; Ananthakrishnan et al. 1968; Krishnamurti and Ramanathan 1982; Ananthakrishnan and Soman 1988; Wang et al. 2001, 2009; Pai and Rajeevan 2009). The mean onset date of summer monsoon rainfall over India has been around 1 June. By the first week of July, the southwest monsoon is established over the entire subcontinent. The southwest monsoon starts withdrawing from the extreme northwestern portion of India by the beginning of September.

Intra-Seasonal Oscillations (ISOs)

The strength of summer monsoon rainfall is modulated by the intra-seasonal variability which is characterized by the active/break spells of enhanced/decreased precipitation over India. The active/break spells over India play an essential role in modulating mean monsoon rainfall (e.g. Ramamurthy 1969; Sikka and Gadgil 1980; Rodwell 1997; Webster et al. 1998; Krishnan et al. 2000; Krishnamurthy and Shukla 2000, 2007, 2008; Annamalai and Slingo 2001; Goswami and Ajayamohan 2001; Lawrence and Webster 2001; De and Mukhopadhyay 2002; Goswami et al. 2003; Waliser et al. 2003; Kripalani et al. 2004; Wang et al. 2005; Mandke et al. 2007; Goswami 2005; Waliser 2006). Active and break episodes, characteristic of subseasonal variations of the ISM, are associated with enhanced (decreased) rainfall over central and western India and decreased (enhanced) rainfall over the southeastern peninsula and eastern India (Singh et al. 1992; Krishnamurthy and Shukla 2000; Krishnan et al. 2000; Goswami et al. 2003).

ISOs are influenced by various factors such as variations in the position and strength of the continental monsoon trough, quasi-periodic oscillations of the monsoon, as well as synoptic systems such as lows and depressions (Shukla 1987 Yasunari 1979; Sikka and Gadgil 1980). These quasi-periodic fluctuations play a major role in determining the amplitude of seasonal mean of individual summer seasons by modulating the strength and duration of active/break

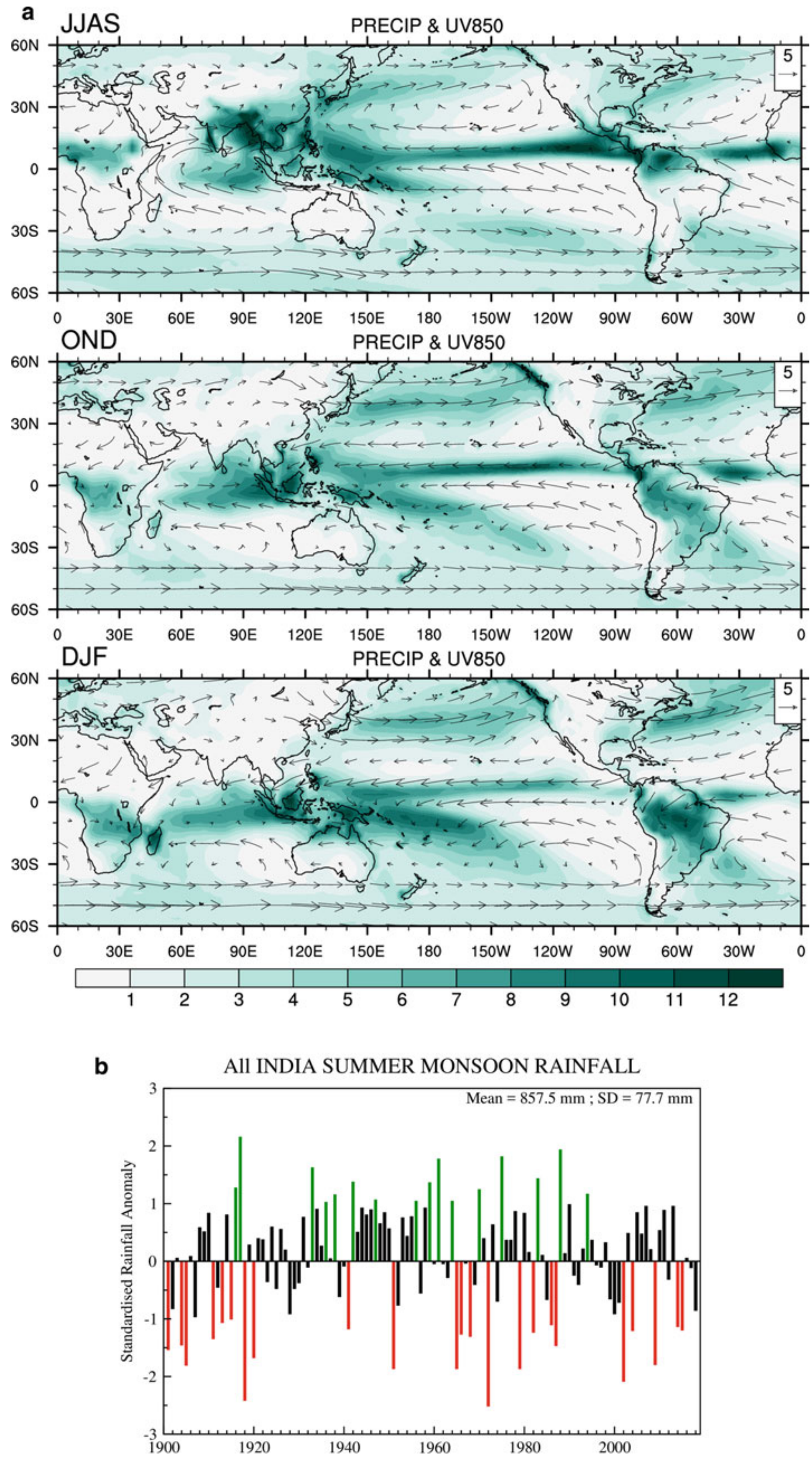

Fig. 3.1 **a** Precipitation (mm/day) and 850 hPa (m/s) wind climatology for southwest monsoon period (JJAS; upper panel), northeast monsoon period (OND; middle panel) and winter monsoon season (DJF; lower panel). GPCP rainfall and ERA-interim winds for the period of 1979–2017 are used. **b** Interannual variability of all-India summer monsoon rainfall (1901–2018). The excess (green), deficit (red) and normal (black) monsoons are also shown. *Data Source*: IMD 0.25^{o} x 0.25^{o} gridded daily rainfall data (Pai et al. 2014)

spells of the ISM through the northward-propagating 30–60-day mode and the westward-propagating 10–20-day mode (e.g. Krishnamurti and Bhalme 1976; Keshavamurty and Sankar Rao 1992). Monsoon intra-seasonal oscillation has a seminal role in influencing the seasonal mean, and its interannual variability (e.g. Goswami and Chakravorty 2017).

Synoptic systems
In the south Asian monsoon region, the synoptic systems are mainly of tropical nature. They play an important role for the onset and advance of the monsoon during June, distribution of rainfall during the peak phase of the monsoon in July and August and withdrawal of the monsoon from mid-September to mid-October. Much of the monsoon rainfall over the central plains of India is associated with the low-pressure systems which develop over the north Bay of Bengal and move onto the subcontinent along a west-northwesterly track (see Chap. 7). The low-frequency Madden–Julian Oscillations (MJO) (Madden and Julian 1972) moving eastward on the 30–40-day scale along near-equatorial belts, on several occasions trigger northward-moving organized convective episodes. These systems sometimes interact with extratropical systems of the northern hemisphere and produce extremely heavy rainfall in some parts of northern India (Pisharoty and Desai 1956; Ramaswamy 1962).

Orographic precipitation
Capacious rainfall rates are generally noticed over the Western Ghats (WG) and north and northeast region of India during the summer monsoon season. These regions have unique characteristics of mountainous terrain that acts as a barrier to southwesterly winds (e.g. Patwardhan and Asnani 2000a, b; Tawde and Singh 2015). The windward side of WG receives highest intense rainfall and leeward side of the WG is a strong rain shadow region. Rain shadow areas differ from one region to another along Karnataka, Maharashtra and Kerala due to the complexity of mountain terrains. Intense orographic rainfall is confined up to 800 m height in the WG (e.g. Rahman et al. 1990). High rainfall spells over the west coast of India are associated with warm Sea Surface Temperature (SST), low-level convergence, high CAPE and low convective inhibition (CIN) (e.g. Maheskumar et al. 2014).

The Himalaya mountain range acts as a barrier by blocking the warm moist monsoon air masses primarily on south-facing slopes and preventing their migration on the other side creating a prominent rain shadow contrast (see Chap. 11). Nearly 80% (20%) of the annual rainfall occur in the Himalayas due to southwest (winter) monsoon. The nature of the convective systems varies dramatically from the western to eastern foothills of Himalayas.

Box 3.2 Precipitation teleconnections with modes of climate variability
ENSO and IOD
The year-to-year variability of Indian monsoon rainfall (Fig. 3.1b) is governed by the slowly varying surface features. El Nino conditions in the Pacific play a major role in modulating the interannual variability of ISM rainfall (Sikka 1980, 1977; Pant and Parthasarathy 1981; Rasmusson and Carpenter 1983; Webster et al. 1998). Almost 50% of the droughts are associated with ENSO (see Chap. 6), however in the last few decades the ENSO-Monsoon relationship has been weakened (e.g. Kripalani and Kulkarni 1997; Krishna Kumar et al. 1999), frequency and intensity of droughts have been increased and some of them are not associated with ENSO.

The coupled mode in the Indian Ocean (Indian Ocean Dipole; IOD; Saji et al. 1999) is also known to modulate interannual variability of ISM rainfall. A positive relationship between IOD and ISM rainfall is well known (Ashok et al. 2004; Saji et al. 1999; Saji and Yamagata 2003). The positive (negative) IOD significantly dilutes the influence of El Nino (La Nina) on the Indian monsoon (Ashok et al. 2004; Chowdary et al. 2015). There are more frequent positive IOD events in recent decades due to the rapid warming of the Indian Ocean (e.g. Cai et al. 2018). In addition to IOD and ENSO, there is a strong link between ISM rainfall and the equatorial Indian Ocean oscillation (EQUINOO; Gadgil et al. 2004). In general, positive phase of the EQUINOO is favourable for a good monsoon. Association between EQUINOO and ENSO also determines the variations in ISM rainfall on the interannual time scale.

Eurasian snow cover
Eurasian snow cover also plays a major role in the year-to-year variability of ISM rainfall (Blanford 1884). Generally, positive Eurasian snow cover anomalies during winter and spring tend to be followed by an anomalous deficit rainfall over the Indian subcontinent in the subsequent summer monsoon season, while negative snow cover anomalies tend to be followed by abundant rainfall (Bhanu Kumar 1987; Bamzai and Kinter 1997). It has been observed that all non-ENSO related droughts over India have been

associated with excessive snow depth over Eurasia (Kripalani and Kulkarni 1999).

Other than ENSO, the Atlantic, western North Pacific circulation changes (e.g. Chowdary et al. 2019; Srinivas et al. 2018) also play a role in monsoon interannual and decadal variability (Sankar et al. 2016; Yadav 2017). The variabilities in the rainfall, as well as the teleconnections of monsoon, could be natural, but there is an intriguing possibility of global warming to modulate these variations. It is suggested that the weakening linkage between ENSO and ISM, despite the increase in ENSO activity, could be due to global warming (Krishna Kumar et al. 1999). Also the anomalous warming over the Eurasian land mass and enhanced moisture conditions over the Indian region in a global warming scenario could have contributed to the weakening of the influence of warm ENSO events on ISM rainfall (e.g. Ashrit et al. 2001). Moreover, the warming of the Indian Ocean at a faster rate than the global oceans (Roxy et al. 2014) has implications on the variability of rainfall over India, by playing a major role in the declining trend of ISM rainfall (Preethi et al. 2017).

Decadal Variations

Variations in ISM rainfall are characterized by distinct epochs typically of about three decades, of above and below normal monsoon activity (e.g. Parthasarathy et al. 1991a, b; Kripalani and Kulkarni 1997). The observational, paleo-climatic and simulated datasets show increased (decreased) ISM rainfall during the positive (negative) phase of the Atlantic Multidecadal Oscillation (AMO) (e.g. Goswami et al. 2006; Joshi and Rai 2015; Krishnamurthy and Krishnamurthy 2015). The leading mode of SSTs in the North Pacific Ocean, Pacific Decadal Oscillation (PDO) with periodicities of 15–25 years and
50–70 years (e.g. Mantua and Hare 2002), could negatively impact ISM rainfall (e.g. Krishnan and sugi 2003; Krishnamurthy and Krishnamurthy 2013). The high correlation between the inter-decadal component of variability of ISM with that of Nino-3 SST highlights the importance of El Nino-Monsoon relationship (Parthasarathy et al. 1994; Kripalani et al. 1997; Kripalani and Kulkarni 1997; Mehta and Lau 1997; Krishnamurthy and Goswami 2000). This indicates that low-frequency modulation of summer monsoon could largely influence rainfall over the Indian subcontinent. Along with this, a strong multi-decadal variability with alternate wet (above normal) and dry (below normal) epochs of monsoon rainfall has been observed in the instrumental records extending back to 150 years (Kripalani and Kulkarni 2001; Joseph et al. 2016; Preethi et al. 2017).

Northeast and Winter Monsoon precipitation

While most parts of India receive the major share of the annual rainfall during southwest/summer monsoon season (from June to September), southeast peninsular India falls under the rain shadow region during this season. During the northeast monsoon season from October to December (OND), the zone of maximum rainfall migrates to southern India and the prevailing winds become northeasterly (Fig. 3.1a; middle panel) (e.g. Ramaswamy 1972; Dhar and Rakhecha 1983; Singh and Sontakke 1999; Balachandran et al. 2006; Rajeevan et al. 2012). The northeast monsoon rainfall shows strong interannual variability (28%), which is more than twice the variability of southwest monsoon rainfall (11%) (e.g. Nageswara Rao 1999; Sreekala et al. 2011). The normal date of the northeast monsoon onset is 20th October with a standard deviation of 7–8 days (Raj 1992).

The relationship between ENSO and northeast monsoon has been strengthened during 1979–2005 (Kumar et al. 2007) while it is weakened in the decade of 2001–2010 (Rajeevan et al. 2012). Local air-sea interaction within the Indian Ocean also modulates the northeast monsoon rainfall (Yadav 2013). IOD-related circulation is found to be an important local forcing mechanism for northeast monsoon (Kripalani and Kumar 2004).

During winter (December to February), cold air masses originating from the Siberian High move southward (Fig. 3.1; bottom panel), and lead to interaction between high northern latitudes and the tropics (Wang et al. 2003). Western disturbances from the Mediterranean to Central Asia transport moisture to the Indian winter monsoon contributing significantly to annual precipitation in the Himalaya region (Dimri 2013; Dimri et al. 2015).

3.2 Observed Changes in Mean Precipitation and Circulation

3.2.1 Precipitation Records in Paleo Time Scale—Inferences from Proxies

The paleoclimate proxy data of monsoonal record of the past 640,000 years suggest that the millennial-scale variability arose by the solar insolation changes which are caused by precession and obliquity (Cheng et al. 2016). In the last 11,000 years (The Holocene age), summer monsoon is declining with variability at the multi-decadal scale to centennial scales (Chao and Chen 2001). The decline of the summer monsoon is linked to, among other factors, the southward migration of the ITCZ as a result of the decrease in solar insolation (Fleitmann et al. 2007). Additionally, prolonged wet/drought periods of multi-decadal and century-scales have occurred during the last 4000 years (Sinha et al. 2011; Prasad et al. 2014) with notable century scale long declining trends 1550–2200 years BP (Roman Warm Period; RWP) and 100–550 years BP (Little Ice Age; LIA) and an increasing trend during 650–1050 year BP (Medieval warm period; MWP) in summer monsoon (Trends marked by arrows; Fig. 3.2). Abrupt changes in monsoon around 2800 years BP and 2350 years BP have been attributed to solar variability (Sinha et al. 2018). Tree ring-based studies from the Himalayan region reveal a declining trend in the summer monsoon over the last 200 years with the possible linkages to large-scale greenhouse warming, and anthropogenic aerosol emissions (Xu et al. 2013; Shi et al. 2017). Speleothem (cave deposits)-based 4000-year-long monsoon reconstructions from central, peninsular and northeast India indicate that the monsoon has undergone multi-decadal changes of larger magnitude in the Holocene than in the last 200 years alone (Fig. 3.2). Therefore, as per the available paleo records, changes in the monsoon due to the regional forcing such as anthropogenic aerosol emissions are difficult to detect against the large multi-decadal natural variability.

3.2.2 Recent Changes in Precipitation

The mean summer monsoon rainfall (JJAS) over India from 1979 to 2005 from multiple observational datasets is shown in Fig. 3.3. It is noted that, in general, all datasets show a high rainfall zone over east-central India and low rainfall zones over northwest India, northern parts of Kashmir, and the rain shadow area of southeast India. The discrepancy in observations occurs mainly over northern parts of India, Himalayan region (e.g., Prakash et al. 2015) and northeast India (Bidyabati et al. 2017).

The annual rainfall averaged over Indian landmass does not show any trend over the period 1901–2015. However, in the recent period 1951–2015 as well as 1986–2015 the annual rainfall series shows decreasing trend (though not statistically significant, or evident in all the datasets). The summer monsoon rainfall series averaged over India landmass does not show any long-term trend on a century-scale where it has been found that the contribution from increasing heavy rain events has been offset by decreasing moderate rain events (Goswami et al. 2006). However, a downward trend of rainfall over the Indian subcontinent has been observed in the period 1951–2004 (Kulkarni 2012). The decreasing tendency of summer precipitation is found

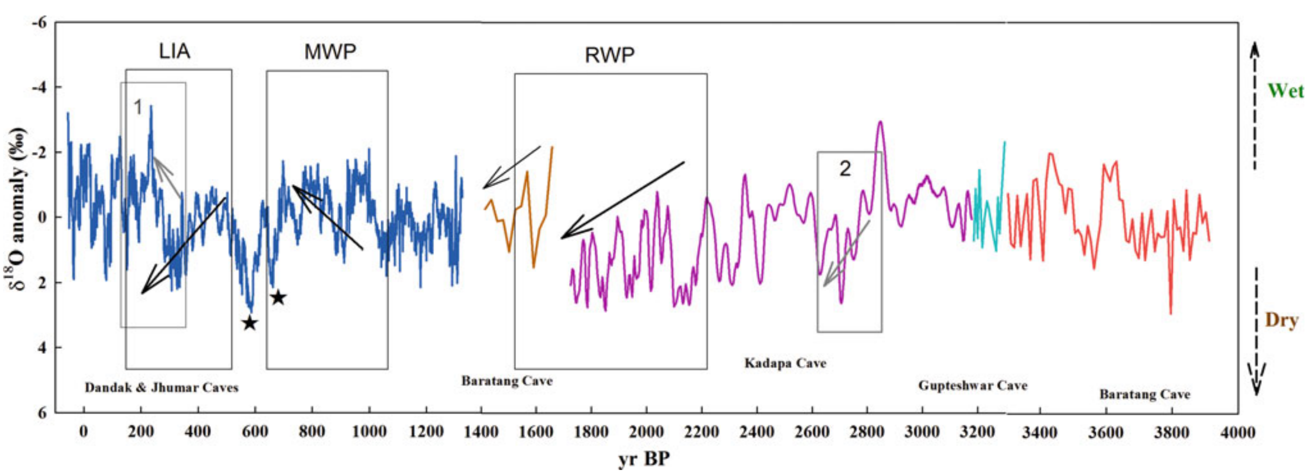

Fig. 3.2 Reconstruction of the ISM of the past 4000 years from a synthesis of records of δ^{18}O of speleothems from Kadapa (Andhra Pradesh), Baratang (Andaman) and Gupteshwar (Orissa) (right *y*-axis) and Dandak-Jhumar (Chhattisgarh -Meghalaya) composite (left *y*-axis). Variations/trends in summer monsoon associated with major climatic events of the past such as LIA, MWP and RWP are highlighted. Abrupt changes in monsoonal strength (box 1 and 2), as well as major drought periods, are also shown (Sinha 2018; marked with stars). Some gaps in the time series are due to the poor resolution of available records for those particular periods, thus, unaccounted

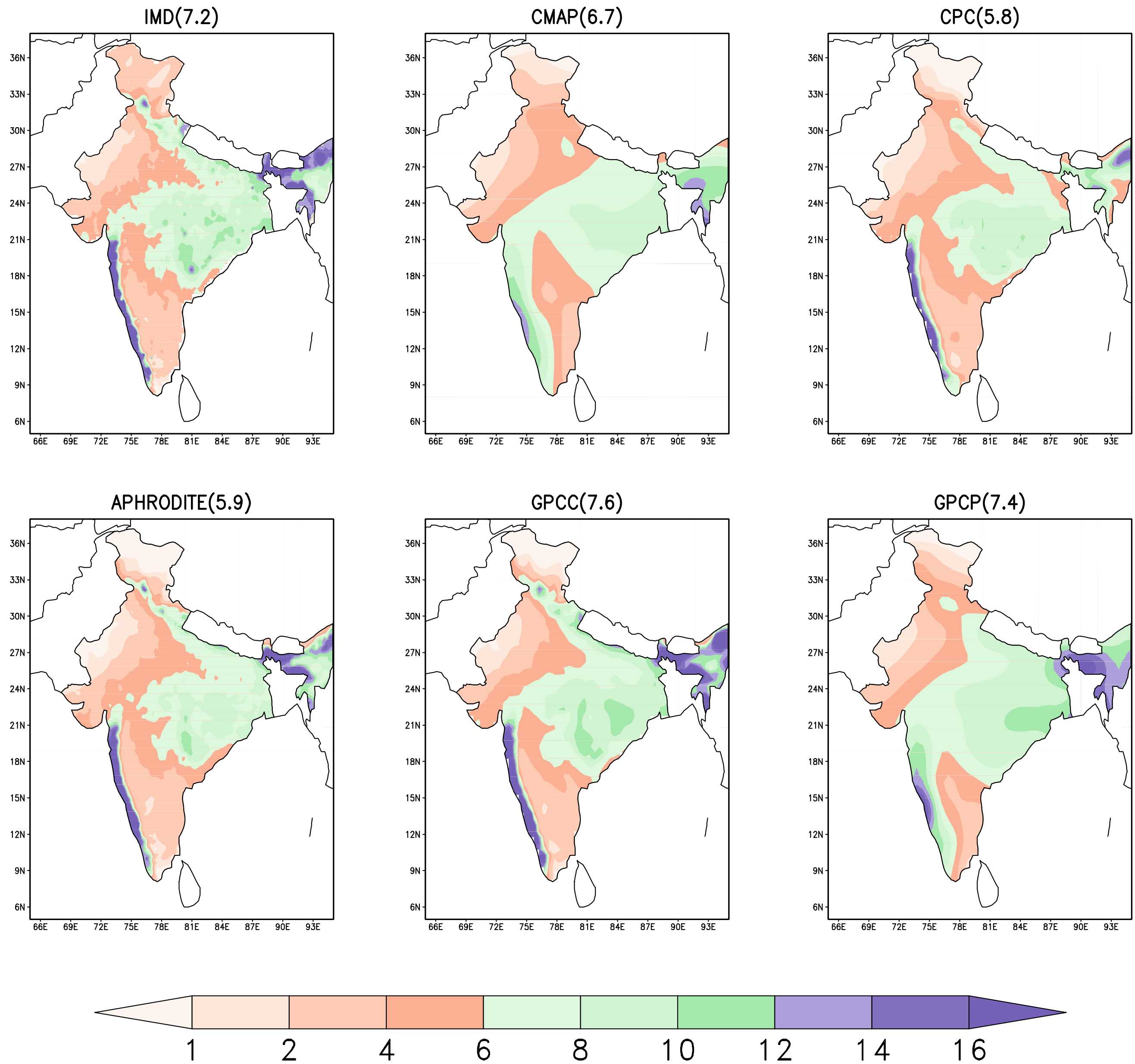

Fig. 3.3 Mean summer monsoon (June through September) rainfall over India from 1979 to 2005 in multiple observed datasets. The dataset and all-India mean rainfall (mm/day) are given in brackets

to have accelerated during 1971–2002 (Kothawale et al. 2008) and is evident in multiple datasets (Annamalai et al. 2013). This secular decline in mean rainfall is attributed to wakening monsoon Hadley circulation. A decreasing trend in rainfall may also be a result of multi-decadal epochal variability associated with an east-west shift in monsoon rainfall due to anomalous warming of the Indo-Pacific warm pool (Annamalai et al. 2013; Guhathakurta et al. 2015) or weakening of the land-ocean temperature gradient (Kulkarni 2012). The drying trend may be attributed to a number of factors such as: (a) increasing anthropogenic aerosol concentration over northern hemisphere which may cool northern hemisphere and slowing of the tropical meridional overturning circulation (e.g. Ramanathan et al. 2005; Chung and Ramanathan 2006; Bollasina et al. 2011, 2014); (b) to increasing trend of the Pacific Decadal Oscillation (Salzmann and Cherian 2015); (c) to significant weakening of monsoon low-level southwesterly winds, the upper-tropospheric tropical easterlies from the out-flow aloft, the large-scale monsoon meridional overturning circulations (Rao et al. 2010; Joseph and Simon 2005; Sathiyamoorthy 2005; Fan et al. 2010; Krishnan et al.

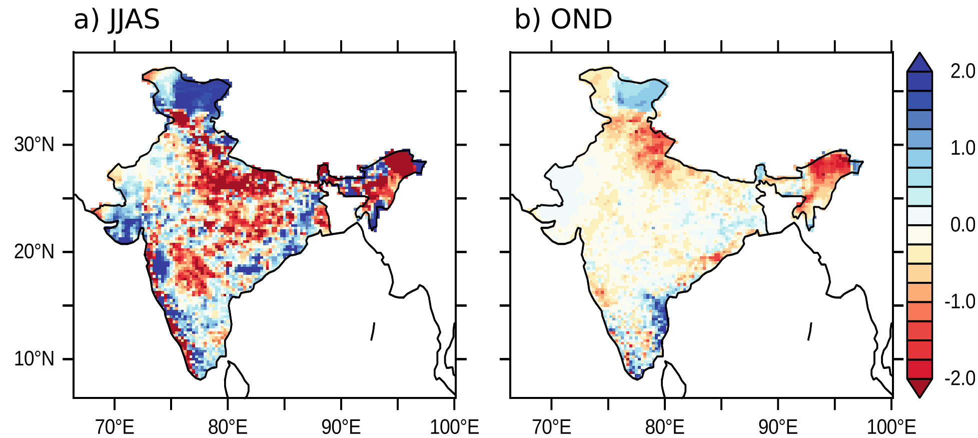

Fig. 3.4 Linear trends (mm/day over 64 years) in the southwest (left) and northeast (right) monsoon rainfall from 1951 to 2015 based on IMD data

2013); (d) to strong Atlantic Multidecadal Oscillation (AMO) which weakens the meridional temperature gradient resulting in early withdrawal of monsoon over India and thus reducing the mean rainfall (Goswami et al. 2006); (e) to significant increase in the duration and frequency of 'monsoon- breaks' (dry spells) over India since the 1970s (e.g. Ramesh Kumar et al. 2009; Turner and Hannachi 2010);(f) to rapid warming of western Indian ocean which reduces the meridional temperature gradient dampening the monsoon circulation (Gnanaseelan et al. 2017; Roxy 2015; Roxy et al. 2015) and (g) to changes in land use/land cover (Niyogi et al. 2010; Pathak et al. 2014; Paul et al. 2016; Krishnan et al. 2016). However, enhanced warming over Indian subcontinent and comparatively slower rate of warming over India Ocean has favoured the land-ocean temperature gradient in the recent decade (2002–2014) and helped a possible short-term revival of the monsoon over India at the rate of 1.34 mm/day/decade (Jin and Wang 2017). Thus, over the recent three decades (1986–2015) all-India summer monsoon shows a decreasing tendency, but the decline is not statistically significant.

There is considerable spatial variability in precipitation changes. As compared to the period 1901–1975, rainfall has reduced by 1–5 mm/day during 1976–2015 over central parts of India (the core monsoon zone), Kerala and extreme northeastern parts and increased over the Jammu and Kashmir region as well as in parts of western India (Kulkarni et al. 2017). Regional anthropogenic forcings such as from aerosols and land-use change from urbanization and agricultural intensification could be dominant contributors to this recent spatial variability (Paul et al. 2018).

Trends in Indian rainfall records have been extensively studied, but the subject remains complicated by the high spatiotemporal variability of rainfall arising from complex atmospheric dynamics and, to some extent, differences that emerge from the methods used in creating the datasets. Monsoon rainfall has shown moderate increasing trends in 27 (out of 36) subdivisions across India (Guhathakurta and Rajeevan 2008). The linear trend in annual as well as seasonal rainfall shows a statistically significant decreasing trend over Jharkhand, Chhattisgarh, and Kerala, and eight subdivisions, viz. Gangetic WB, West UP, Jammu and Kashmir, Konkan and Goa, Madhya Maharashtra, Rayalaseema, Coastal AP and North Interior Karnataka show increasing trends (Guhathakurta and Rajeevan 2008). Based on high-resolution gridded data for 1901–2015, there are statistically significant decreasing trends in annual as well as seasonal rainfall over Kerala, Western Ghats and some parts of central India including Uttar Pradesh, Madhya Pradesh, and Chhattisgarh as well as some parts of the northeastern states. Whereas rainfall over Gujarat, Konkan coast, Goa, Jammu and Kashmir and east coast shows a significant increasing trend (Fig. 3.4).

Climate change is not just affecting the southwest monsoon but is also driving changes in the northeastern monsoon. The variability of northeast monsoon rainfall has increased in the period 1959–2016. Seasonal rainfall has increased over Tamil Nadu, Rayalaseema, as well as south peninsular India because of an increase in the number of high-intensity rainfall events in the recent period compared to 1901–1958 (Nageswararao et al. 2019). Table 3.1 gives summary statistics for rainfall over India based on 1951-2015.

3.2.3 Understanding the Observed Changes in the Summer Monsoon Precipitation

3.2.3.1 Anthropogenic Causes of Observed Precipitation Changes

In general, we can summarize that the unprecedented increase in atmospheric greenhouse gases (GHGs) is responsible for the global rise in temperature, which as feedback to atmospheric dynamics and convection has also led to changes in rainfall characteristics globally (Alexander 2016). Some regional forcings, such as aerosols and land-cover changes, have additionally detectable and notable impact on the monsoon rainfall changes.

Anthropogenic aerosols modulate regional precipitation patterns, over monsoon regions (Bollasina et al. 2011; Krishnan et al. 2016; Undorf et al. 2018). Aerosols play an important role in the earth-atmosphere system through their interactions with solar radiation, clouds and the cryosphere aerosol solar absorption over the Indian monsoon region has a potential role in influencing the monsoon circulation and rainfall distribution (Chap. 5 *provides a summary*). Observed patterns of regional changes in precipitation are missing from the CMIP5 (Coupled Model Intercomparison Project 5) assessments–primarily due to the coarse resolution of models and also due to missing local features that can be important for such regional variability. The current generation of coupled models shows very substantial dry bias in simulating Indian monsoon precipitation over the core monsoon zones of central India, and the Western Ghats. Sabin et al. (2013) used a variable resolution global atmospheric model with telescopic zooming over south Asia (∼35 km) and demonstrated that the high resolution provides particular value addition in simulating better monsoon rainfall over the Indian region. Using the same set of model analysis, recent changes in observed mean monsoon over India (1951–2005) have been attributed to a combined effect of anthropogenic aerosol, equatorial Indian Ocean warming and land-use/land-cover change (Krishnan et al. 2016, Fig. 3.5).

3.2.3.2 Changes in Circulation Features

The core of the Tropical Easterly Jet (TEJ) has been shrinking over the South Asian region (Pattanaik and Satyan 2000). The strength of TEJ has been found to have been decreasing before 2000 (Sathiyamoorthy 2005), but since 2000 has increased at the rate of 1 m/s per year (Roja Raman et al. 2009; Venkat Ratnam et al. 2013). The weakening of the TEJ may be attributed to increase in convection due to the excessive warming of Indian Ocean SST (Joseph and Sabin 2008), cooling of upper-tropospheric temperature over the Tibetan anticyclone region, and a significant warming over the equatorial Indian Ocean which might have resulted in decreasing trend of the upper-tropospheric meridional temperature gradient. These changes have caused a reduction in the strength of the easterly thermal wind at the core region of the TEJ, after the weakening of the TEJ. Further, the weakening of TEJ and associated decrease of easterly shear is attributed to the reduced north-south temperature gradient between the equator and 20°N for the longitude belt of 40°E–100°E, that is, the air temperature on the equator side is increasing compared to the north. These variations are particularly high above 500 hPa (Rai and Dimri 2017).

In response to the global warming, the intensity of the summer monsoon overturning circulation (monsoon Hadley cell) and the associated southwesterly monsoon flow (LLJ) have significantly weakened from the 1950s (Joseph and Simon 2005; Krishnan et al. 2013). An ultra-high-resolution global general circulation model (about 20 km resolution) also shows a stabilization (weakening) of the summer monsoon Hadley-type circulation in response to global warming which has resulted in a weakened large-scale monsoon flow (Rajendran et al. 2012; Krishnan et al. 2013). The weakening of Asian monsoon circulation (Fig. 3.5) may be due to relatively smaller warming in Asia compared to the surrounding regions which make the landmass a 'heat sink' (Zuo et al. 2012). Indeed, the tropospheric temperature over Asia has lowered in recent decades. As a consequence, the meridional and zonal land-sea thermal contrasts are reduced, and the Asian summer monsoon becomes weaker.

3.2.3.3 Observed Changes in Active/Break Spells

The seasonal monsoon strength is mainly modulated by the intra-seasonal variability of the summer monsoon rainfall

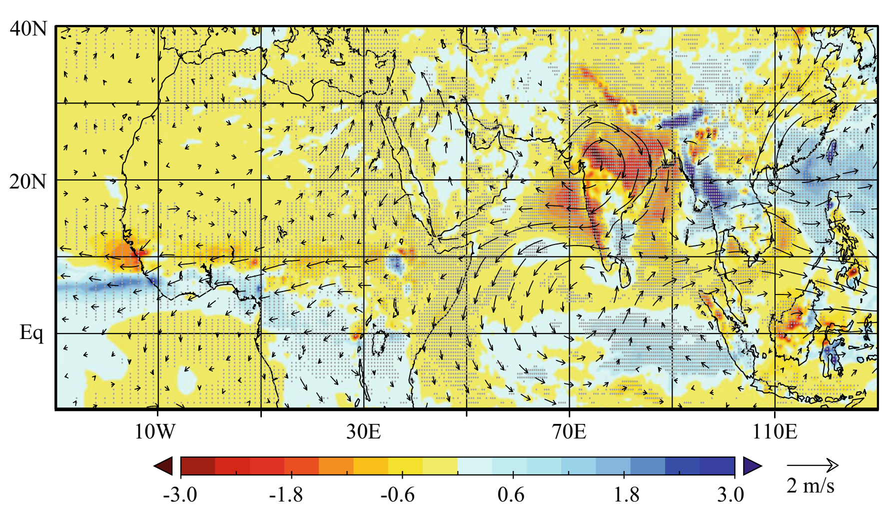

Fig. 3.5 Attribution of the decline in monsoonal rainfall: The difference in mean precipitation (JJAS; mm/day) and low-level circulation at 850 hPa (m/s) between the Historical and Historical natural simulations for the period (1951–2005) from a high-resolution simulation. Regions with a significance level above 95% level are shown with grey dots. Reprinted with permission from Krishnan et al. 2016

which is characterized by the active/break spells of enhanced and decreased precipitation over India (e.g. Ramamurthy 1969; Sikka and Gadgil 1980; Rodwell 1997; Webster et al. 1998; Krishnan et al. 2000; Krishnamurthy and Shukla 2000, 2007, 2008; Annamalai and Slingo 2001; Goswami and Ajayamohan 2001; Lawrence and Webster 2001; De and Mukhopadhyay 2002; Goswami et al. 2003; Waliser et al. 2003; Kripalani et al. 2004; Wang et al. 2005; Mandke et al. 2007; Goswami 2005 and Waliser 2006). The relative strength of the northward-propagating low-frequency (20–60 days) modes has a significant decreasing trend during 1951–2010, possibly due to the weakening of large-scale circulation in the region during the monsoon season. This reduction is compensated by a gain in synoptic-scale (3–9 days) variability. The decrease in low-frequency ISO variability is associated with a significant decreasing trend in the percentage of extreme events during the active phase of the monsoon. However, this decrease is likely balanced by significant increasing trends in the percentage of extreme events in the break and transition phases. These changes are accompanied by a weakening of low-frequency variability (Karmakar et al. 2015, 2017). Also, while there is no change in the distribution of the break events, the number of active spells shows an increase of about 12% in the period 1951–2010; the increase was mainly in the short duration (3–6 days) spells (Pai et al. 2016). A statistically significant increase in the frequency of dry spells (27% higher during 1981–2011 relative to 1951–1980) and intensity of wet spells and statistically significant decreases in the intensity of dry spells have been observed in recent six decades (Singh et al. 2014). The changes in frequency, intensity and speed of intra-seasonal oscillations have been attributed to Indian ocean warming (Sabeerali et al. 2015); developing and decaying phase of ENSO (Pillai and Chowdary 2016); increase in convective available potential energy, low-level moisture convergence and changes in large-scale circulation in upper atmosphere (Singh et al. 2014).

3.2.3.4 Changes in Onset Characteristics

The onset of summer monsoon over India is characterized by the dramatic rise in mean daily rainfall over Kerala (Ananthkrishnan and Soman 1988; Soman and Kumar 1993). The onset of the ISM has been defined with various dynamic (Koteswaram 1958; Ananthakrishnan et al. 1968; Krishnamurti and Ramanathan 1982; Wang et al. 2001; 2009; Pai and Rajeevan 2009) and thermodynamic indices (Ananthakrishnan and Soman 1988; Fasullo and Webster 2003; Janowiak and Xie 2003). Objective definitions of South Asian summer monsoon onset include measures such as the increase of rainfall above a threshold (Wang and Lin 2002), transition in vertically integrated moisture transport (Fasullo and Webster 2003), reversal of surface wind (Ramage 1971), and intensification of the lower level Somali jet (Taniguchi and Koike 2006; Wang et al. 2009). As per these different definitions, the mean onset date of summer monsoon rainfall over India has been stable around 1 June. In recent decades, the monsoon onset over India is seen to be delayed to 5th June since 1976 (Sahana et al. 2015), which can be attributed to the net decrease in moisture supply from the Arabian Sea in the post-1976 period. The interannual variability of the onset date is associated with ENSO with early onsets preceded by La Nina, and late onsets preceded by El Nino (e.g. Noska and Mishra 2016).

3.3 Projected Changes in Precipitation Over India

Understanding the projected future changes in precipitation has a profound importance for policy. In this report, the assessment of rainfall changes over India is carried out based on the multiple ensemble member simulations from CMIP5, CORDEX-SA (COordinated Regional Downscaling EXperiment-South Asia) and NEX-GDDP (Nasa earth Exchange-Global Daily Downscaled Products) in which CMIP5 is the parent GCM, CORDEX is dynamically downscaled to 50 km × 50 km grid resolution, and NEX-GDDP is statistically downscaled to 25 km × 25 km grid resolution.

Historical and projected changes until the end of the twenty-first century based on various simulations (28 from CMIP5, 16 from CORDEX and 19 from NEX; see the list of models in Tables 3.2 and 3.3) are provided in this section. The future changes are mostly quantified as percentage changes in the near future (2040–2069) and far future (2070–2099) epochs. We provide our analysis for annual, summer (JJAS) and winter (OND) seasons in all cases. Mostly the analysis is restricted to the Indian landmass, by masking out the seas, and regions outside the geographical area of India. Projections are stated with respect to the standard reference period of 1976–2005.

Mean precipitation from multi-model ensemble simulations for annual, JJAS and OND seasons is shown in Fig. 3.6.

The change in mean precipitation over India for the annual, summer and winter seasons is presented as box-whiskers in Fig. 3.7. A comparison of the various sources of climate data used in this assessment shows a consistent enhancement in precipitation across the Indian landmass throughout the twenty-first century. The box-whiskers also highlight the spread among the three suites of experiments. The variability is comparatively high during the winter monsoon season (OND). Comparing with the coarse resolution CMIP5 simulation, the high-resolution CORDEX and NEX simulations show higher variability irrespective of seasons. This increased variability in the

Table 3.1 Statistics of rainfall over the Indian landmass

	Monsoon	Post-monsoon	Annual
Seasons	JJAS	OND	
Mean (mm)	858.0	119.1	1142.1
Std Dev	80.2	29.5	98.8
% Contribution to annual average	75.1	10.4	
Max RF (Year)	1011.7 (1988)	205.1 (1956)	1359.6 (1990)
Min RF (Year)	665.2 (1972)	63.5 (2011)	922.4 (1972)

Table 3.2 List of CORDEX South Asia simulations used

CORDEX -SA	RCM used	Contributing RCM modelling centre	Driving CMIP5 model	CMIP5 modelling centre
IITM-RegCM4	ICTP, regional climate model version 4 (Giorgie et al. 2012)	CCCR, IITM, India	CCCma-CanESM2	*Canadian Centre for Climate Modelling and Analysis (CCCma), Canada*
			NOAA-GFDL-ESM2M	National Oceanic and Atmospheric Administration-Geophysical Fluid Dynamics Laboratory, USA
			CNRM-CM5	Centre National de Recherches Météorologiques, France
			MPI-ESM-MR	Max Planck Institute for Meteorology, Germany
			IPSL-CM5A-LR	The Institute Pierre Simon Laplace, France
			CSIRO-Mk.6	Commonwealth Scientific and Industrial Research Organisation, Australia
SMHI-RCA4	Rossby Centre regional atmospheric model version 4 (Samuelsson et al. 2011)	Rossby Centre, Swedish Meteorological and Hydrological Institute, Sweden	ICHEC-EC-EARTH	Irish Centre for High End Computing, European Consortium
			MIROC-MIROC5	Model for Interdisciplinary Research On Climate, Japan Agency for Marine-Earth Science and Technology, Japan
			NCC NorESM1	Norwegian Climate Centre, Norway
			MOHC-HadGEMEM2-ES	Met Office Hadley Centre for Climate Change, UK
			CCCma-CanESM2	*Canadian Centre for Climate Modelling and Analysis (CCCma), Canada*
			NOAA-GFDL-ESM2m	National Oceanic and Atmospheric Administration-Geophysical Fluid Dynamics Laboratory, USA
			CNRM-CM5	Centre National de Recherches Météorologiques, France
			MPI-CM5A-MR	Max Planck Institute for Meteorology, Germany
			IPSL-CM5A-MR	The Institute Pierre Simon Laplace, France
			CSIRO-Mk.6	Commonwealth Scientific and Industrial Research Organisation, Australia

Table 3.3 List of CMIP5 and NEX Models used in this study. The availability of NEX-GDDP statistically downscaled product for the respective model is depicted with "†"

CMIP5	NEX	CMIP5 modelling centre
ACCESS1-0	†	Australian Community Climate and Earth System Simulator, Australia
BNU-ESM	†	Beijing Normal University Earth System Model, China
BCC-CSM1-1	†	Beijing Climate Centre Climate System Model, China
CCSM4 CESM1-BGC	† †	Community Climate System Model, NCAR, USA
CNRM-CM5	†	Meteo-France/Centre National de Recherches Meteorologiques, France
CanESM2	†	*Canadian Centre for Climate Modelling and Analysis (CCCma), Canada*
CMCC-CM		Centro Euro-Mediterraneo sui Cambiamenti Climatici, Italy
CSIRO-Mk3-6-0	†	Commonwealth Scientific and Industrial Research Organisation (CSIRO), Australia
GFDL-ESM2M GFDL-ESM2G GFDL-CM3	† † †	Geophysical Fluid Dynamics Laboratory, National Oceanic and Atmospheric Administration (NOAA), USA
GISS-E2H GISS-E2-R		National Aeronautics and Space Administration (NASA)/Goddard Institute for Space Studies (GISS), USA
HadCM3 HadGEM2-ES HadGEM2 CC		Met Office Hadley Centre, UK
INMCM4	†	Institute for Numerical Mathematics, Russia
IPSL-CM5A-MR IPSL-CM5A-LR	† †	Institute Pierre Simon Laplace, France
MIROC5 MIROC-ESM-CHEM MIROC-ESM	†	Centre for Climate System Research (University of Tokyo), National Institute for Environmental Studies and Frontier Research Center for Global Change (JAMSTEC), Japan
MPI-ESM-LR MPI-ESM-MR MRI-CGCM3	† † †	Max Planck Institute for Meteorology, Germany
MRI-CGCM3		Meteorological Research Institute, Japan
NorESM1-M	†	Norwegian Climate Centre, Norway

future climate is noted in all three experiments, especially for the RCP8.5 scenario as noted by Singh and Achutarao 2018. The quantitative estimate of future changes in annual mean precipitation from different Reliability Ensemble Average (REA) for projected change in precipitation (mm/day) along with uncertainty range is summarized in Table 3.4.

The percentage change in projected mean precipitation pattern for near and far future from RCP4.5 and RCP 8.5 is shown in Fig. 3.8. Multi-model mean change suggests wetter condition over India in near and far future on average. Slightly different scenario is projected in the CORDEX simulations over the northwest Indian region with a 10% drier condition than its present-day mean for near future in RCP4.5 simulations. During the winter months, northeast India is projected to witness a moderate deficit condition in the near future (both in CMIP5 and NEX); in addition to this, CORDEX models suggest a potential reduction over the Himachal and Jammu belt. From the extreme scenario (RCP8.5, both near and far future) and far future in RCP4.5, consistent pattern emerges among the three sources, irrespective of variations in their spatial resolution and methodologies followed. The changes in annual mean precipitation surpass 10% above baseline over the west coast and southern locale of the Indian landmass in the RCP4.5 scenario in the near future, and exceed 20% in far future (Figs. 3.8 and 3.9). Over the rest of India, the precipitation changes are not significant for the near future up to the mid-twenty-first century, yet in the long-term; increment surpasses 10% over northwest and the adjoining territory of the nation. The long-term projected annual precipitation increment surpasses 10% over most parts the Indian landmass.

3.3.1 Future Changes in the Summer Monsoon

ENSO and IOD typically exert an offsetting impact on Indian summer monsoon rainfall (ISMR), with an El Niño event tending to lower, whereas a positive IOD tending to increase ISM rainfall (Ashok and Saji 2007). In a recent study, Li et al. (2017) showed that CMIP5 models simulate an unrealistic present-day IOD-ISMR correlation due to an

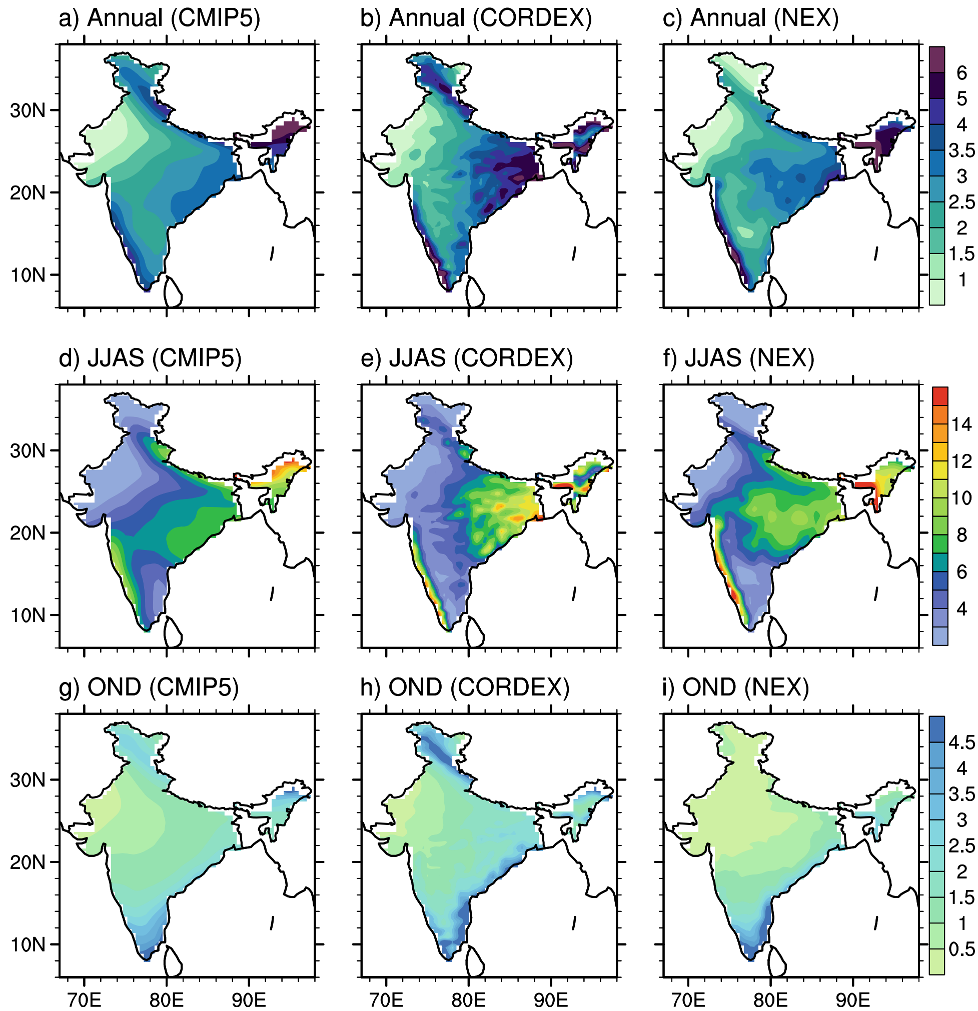

Fig. 3.6 Mean precipitation (mm/day, 1976–2005) from multi-model ensemble simulations for annual, JJAS, and OND seasons from CMIP5, CORDEX-SA, and NEX-GDDP experiments

overly strong control by ENSO. As such, a positive IOD is associated with an ISM rainfall reduction in the simulated present-day climate. They further highlight that the uncertainties in ISM rainfall projection can be in part due to the present-day simulation of ENSO, the IOD, their relationship, and their rainfall correlations. Thus, the natural variability also plays a dominant role in the diverse ENSO-monsoon relationship during the twentieth century (Li and Ting 2015). From CMIP5 models, the analysis by Li et al. (2017) further showed that in future, the enhance SST warming could likely lead to a weak ENSO-monsoon relation as well.

Multi-model average changes considered in the different model sources, in general, suggest wetter future conditions. CMIP5 models project an increase of 6% (RCP4.5) and 8% (RCP8.5) in the near future over the central Indian region (core monsoon zone defined by Rajeevan et al. 2008). Projected changes in rainfall by the end of the twenty-first century are 10% (RCP4.5) and 14% (RCP8.5), respectively. The vast majority of the CMIP models shows enhanced monsoon precipitation due to global warming (e.g. Kitoh et al. 1997; Douville et al. 2000; Ueda et al. 2006; Cherchi et al. 2011; Rajendran et al. 2012; Krishnan et al. 2013); however, they indicate a likely weakening of large-scale monsoonal circulation (Krishnan et al. 2016). Many studies noted that the poor skill in simulating monsoon amplifies the ambiguities in understanding the future changes in projected monsoon rainfall (e.g. Chaturvedi et al. 2012; Saha et al. 2014; Sharmila et al. 2015; Krishnan et al. 2016). Studies have highlighted the wide inter-model spread in the simulated precipitation changes over South Asia, which therefore makes the assessment of regional hydroclimatic response a bit ambiguous in reality (e.g. Kripalani et al. 2007; Annamalai et al. 2007; Turner and Slingo 2009; Sabade et al. 2011; Fan et al. 2010; Hasson et al. 2013; Saha et al. 2014).

Applying the Clausius–Clapeyron equation, the water vapour holding capacity of the atmosphere is expected to increase by about 7% per degree of warming. The enhanced availability of moisture can naturally lead to more precipitation over different parts of the globe (Trenberth 1998; Meehl et al. 2005). Indeed, there is a considerable multiscale feedback through large-scale circulation and various

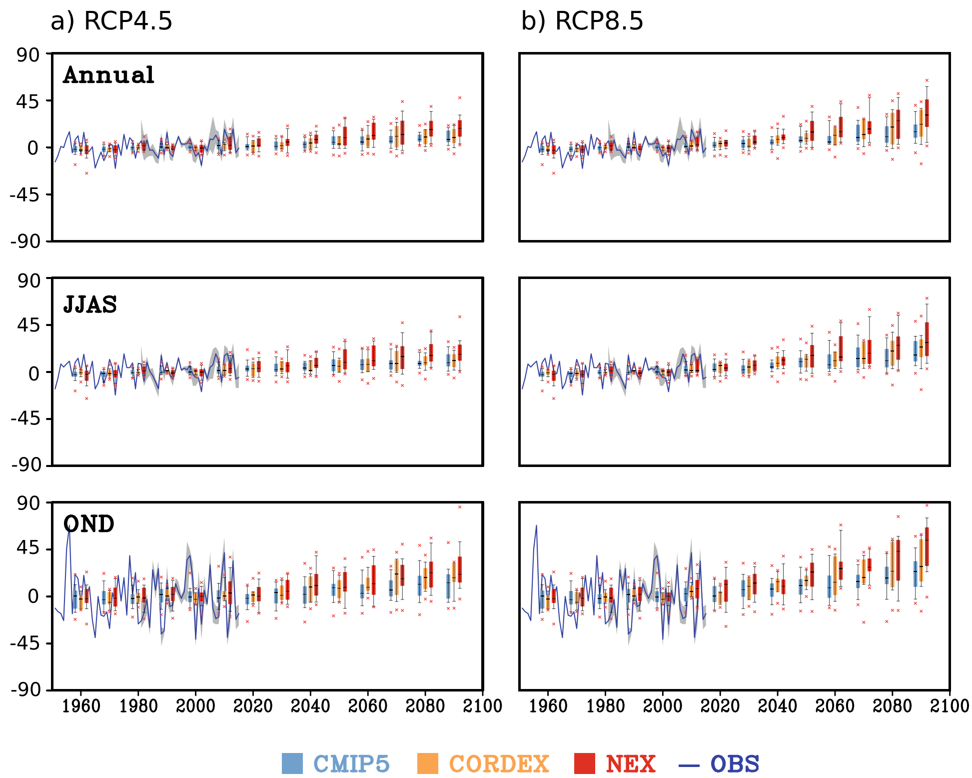

Fig. 3.7 Percentage changes in mean precipitation over the Indian land across CMIP5, CORDEX-SA, and NEX-GDDP simulations from the **a)** RCP4.5 **b)** RCP8.5 scenario for annual, JJAS and OND seasons. Changes in 10-year means, with respect to the reference period (1976–2005), are shown as box-whiskers

Table 3.4 Reliability ensemble average estimate of projected change in mean precipitation (mm/day) over Indian land associated with uncertainty range

	Scenario	Experiments	Annual
Near future	RCP 4.5	CMIP5	0.12 ± 0.11 *(91%)*
		CORDEX	0.16 ± 0.19 *(118%)*
		NEX	0.20 ± 0.21 *(105%)*
	RCP 8.5	CMIP5	0.20 ± 0.17 *(85%)*
		CORDEX	0.28 ± 0.18 *(64%)*
		NEX	0.38 ± 0.20 *(52%)*
Far future	RCP 4.5	CMIP5	0.25 ± 0.18 *(72%)*
		CORDEX	0.32 ± 0.22 *(68%)*
		NEX	0.40 ± 0.23 *(57%)*
	RCP 8.5	CMIP5	0.45 ± 0.21 *(46%)*
		CORDEX	0.58 ± 0.32 *(55%)*
		NEX	0.63 ± 0.31 *(49%)*

thermodynamic and dynamic conditions locally that also contribute to rainfall occurrence. In general, climate simulations hint that the global warming is expected to enhance the Indian summer monsoon precipitation by 5–10%, albeit some climate models suggest less. In the Indian monsoon region, the seasonal movement of the ITCZ brings rainfall over land, and the convection is strongly coupled with large-scale circulation (Goswami 2006; Joseph and Sabin 2008; Gadgil 2018). However, this convective-coupling is relatively weak in most of the CMIP5 models and mostly simulating an increase in precipitation irrespective of the anomalous decrease in monsoon circulation (Krishnan et al. 2016). Sabeerali et al. (2015) postulated that the increase in future ISMR simulated in CMIP5 models is a result of unrealistic local convective precipitation enhancement that is not related to large-scale monsoon dynamics, which is mainly due to the unrealistic representation of stratiform and convective cloud ratio in the coupled model. Using CMIP5 outputs, Sabeerali and Ajaymohan (2018) showed that there is a possibility of a shorter rainy season (defined using tropospheric temperature gradient as outlined by Goswami and Xavier (2005)) by the end of the twenty-first century in the RCP8.5 scenario.

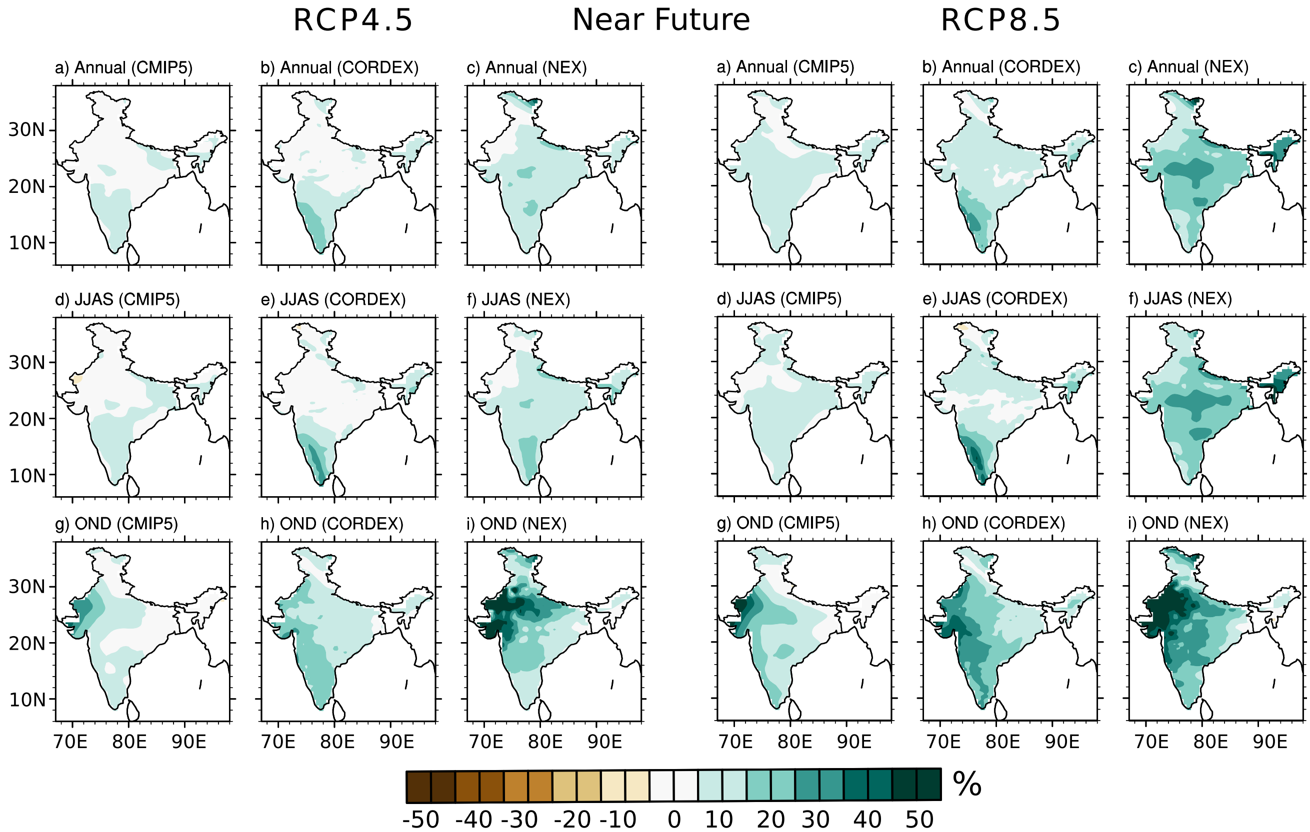

Fig. 3.8 Multi-Model Ensemble (MME) change (%) in annual, JJAS and OND rainfall as projected by CMIP5, CORDEX-SA and NEX-GDDP models for near future with respect to 1976–2005 from RCP4.5 and RCP8.5 scenario

Even though there is ambiguity in changes in projected seasonal mean precipitation, future projections from CMIP5 models show a significant increase in interannual variation during the summer monsoon season (Menon et al. 2013). Studies such as Sarita and Rajeevan (2016) have suggested that the periodicity of El Ninos is likely to shift to a shorter period (2.5–3 year) compared to present day (3–5 year) which will have serious implication on monsoon in inter-annual timescale. The increase in summer monsoon precipitation over South Asian region projected by CMIP5 models is mostly sustained by an increase in moisture supply due to enhanced warming (Mei et al. 2015). By analysing 14 CMIP models, they further showed that towards the end of the century the precipitable water over Indian landmass may increase by 8–16 mm/day, the evapotranspiration by around 0.6 mm/day while the change in moisture convergence around 2.4 mm/day or more under RCP8.5 scenario. Figure 3.10 shows the change in vertically integrated moisture convergence towards the end of the century under RCP4.5 and RCP8.5.

3.3.2 Future Changes in Northeast Monsoon

Interannual variations in the northeast monsoon are much larger compared to the southwest monsoons (Fig. 3.7). Projected changes for the near future and far future from CORDEX, NEX, and CMIP5 show a moderate increase in rainfall over the Indian landmass. Note that significant change is seen only over the region where northeast monsoon is not much significant (Figs. 3.8 and 3.9). Over major areas in Tamil Nadu and coastal Andhra, only modest changes are projected for both RCP 4.5 and RCP 8.5. Parvathi et al. (2017) reported that CMIP5 models project a robust reduction of the wind intensity during the northeast monsoon season especially over the Arabian Sea by the end of the twenty-first century (a reduction of 3.5% for RCP4.5 and 6.5% for RCP8.5, on an average). Despite a decrease in the winter monsoon winds, they noted an increased rainfall (10 ± 2%) in the winter monsoon rain zones in the equatorial Indian Ocean. Over Indian landmass, the precipitation enhancement is minimal in the near future in the RCP4.5

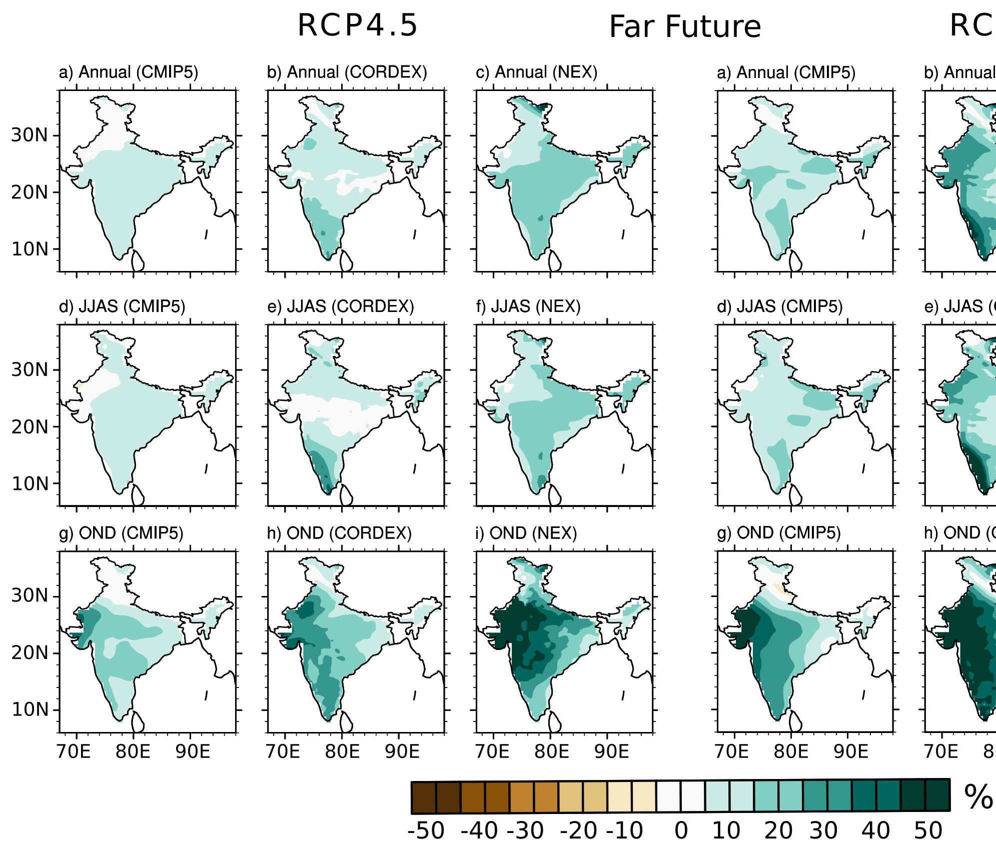

Fig. 3.9 Multi-Model Ensemble (MME) change (%) in annual, JJAS and OND rainfall as projected by CMIP5, CORDEX-SA and NEX-GDDP models in far future with respect to 1976–2005 from RCP4.5 and RCP8.5 scenario

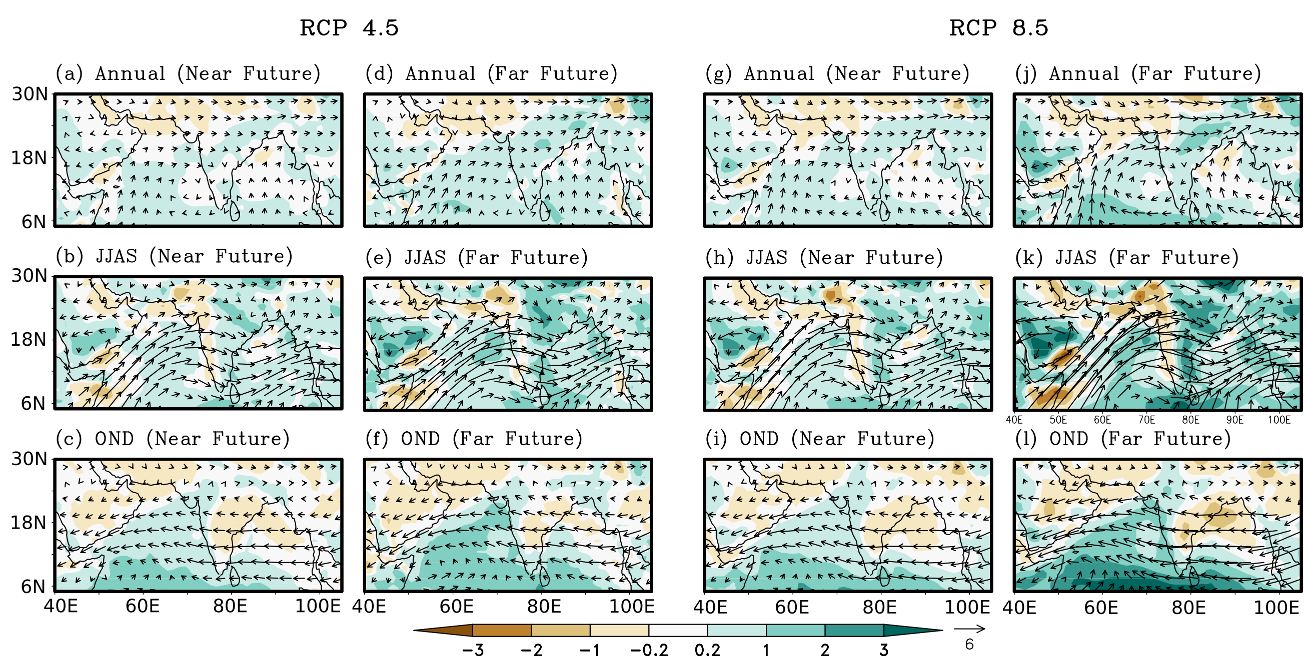

Fig. 3.10 Changes in vertically integrated moisture transport (vectors; Kg m^{-1} s^{-1}) and moisture convergence (shadings; 10^{-6} kg m^{-2} s^{-1}) from surface pressure level to 300 hPa for (**a**, **d**) Annual, (**b**, **e**) JJAS and (**c**, **f**) OND seasons, projected by multi-model ensemble of CMIP5 models for near future (2040–2069) and far future (2070–2099) for RCP4.5and RCP8.5 scenarios, with respect to the reference period (1976–2005)

scenario; however, both projections (RCP4.5 and RCP8.5) shows an increase in precipitation [CMIP5 (10–20%), CORDEX (10–25%) and NEX (15–35%)] by the end of the twenty-first century.

3.4 Changes in Daily Precipitation Extremes

3.4.1 Observed Changes and Their Attribution

Detecting changes in the characteristics of extreme rain events is an important issue in view of their large impacts on human society (Ghosh et al. 2016). It is difficult to attribute a specific extreme event during the monsoon is owing to anthropogenic climate change—like the Uttarakhand surges of 2015, or recent flood in Kerala during 2018 monsoon—yet it is robustly anticipated that a warming atmosphere will result in more severe weather. It has been observed that from 1950 onwards there has been a significant rising trend in the frequency and intensity of extreme heavy rain events over central India, along with a decreasing trend in the moderate rain events (Goswami et al. 2006; Dash et al. 2009; Kulkarni et al. 2017; Krishnan et al. 2016; Roxy et al. 2017) (Fig. 3.11a). Consecutive dry days with minimum spell length of 5 days show significant increase of about 4 days in the period 1951–2015, while consecutive wet days show decrease of about 10 days in this period. Prolonged break spells appear to be more frequent in 1951–2015. Roxy et al. (2016) showed that the widespread changes in extreme rain events are mainly dominated by dynamic response of the atmosphere rather than thermodynamic factor alone. Krishnan et al. (2016) showed that, the enhancement of such deep localized convection, leading to heavy rainfall events, are more likely to happen in an atmosphere with weak vertical shear (Romatschke and Houze 2011). Increased variability of low-level monsoon westerlies (Mishra et al. 2018; Roxy et al. 2017) and warming of north Arabian sea lead to increased moisture supply and thus enhance such events (Roxy et al. 2017). By examining the changes in the distribution of moderate and heavy monsoon precipitation in Historical and GHG, only simulations Krishnan et al. (2016) have shown that along with increase of atmospheric moisture the decrease of easterly vertical shear of the SAM circulation is also pivotal for favouring localized heavy rainfall over the Indian region (Fig. 3.11b). In a recent study, Singh et al. (2014) found statistically significant increase in the intensity and frequency of extreme wet and dry spells during the ISM during the 1951–2011 period.

3.4.2 Future Projections of Precipitation Extremes

The IPCC Special Report on Extremes (SREX; Intergovernmental Panel on Climate Change 2012) appraisal reported that extreme precipitation events globally are certain to rise in the future. From 1950 onwards, the number of extreme precipitation events over Indian landmass has also become more significant than it before (Sillmann et al. 2013; Goswami et al. 2006; Rao et al. 2014). A recent study by Mukherjee et al. (2017) showed that 1–5-day precipitation maxima at 5–500 year return period will increase (10–30%) with anthropogenic warming in RCP8.5 scenario. They further showed that the frequency of precipitation extremes is projected to rise more prominently in the RCP8.5 scenario over southern and central India by the middle and end of the twenty-first century. The analyses of select precipitation-based indices, defined by the Expert Team on Climate

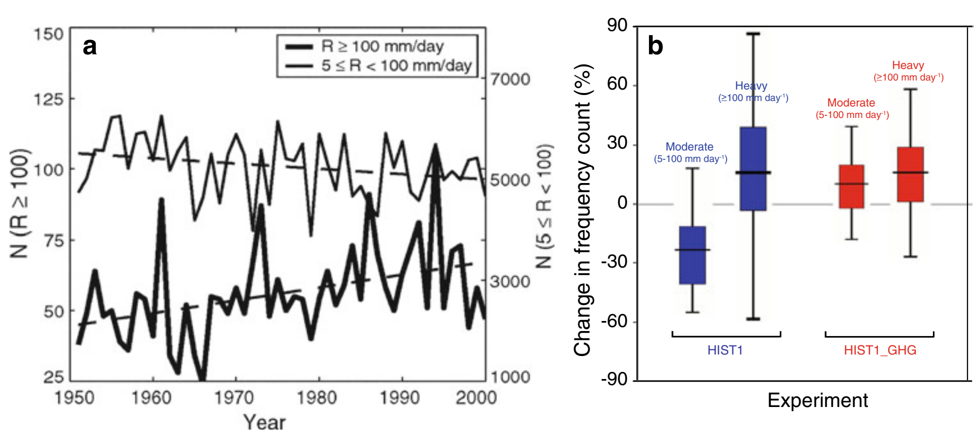

Fig. 3.11 Observed frequency of **a** heavy ($R \geq 100$ mm/day, bold line) and moderate ($5 \leq R < 100$ mm/day, thin line) daily rain events (Goswami et al. 2006). **b** Attribution of changes in moderate and heavy precipitation. *Box-whisker* plot of percentage distributions of yearly count of moderate (5–100 mm day^{-1}) and heavy ($\geq$ 100 mm day^{-1}) events over Central India (74.5°–86.5° E, 16.5°–26.5° N) during the period (1951–2000) from Historical and GHG experiments with respect to the natural only simulation (Krishnan et al. 2016). Permission taken from American Association for the Advancement of Science for Fig. 3.11a.

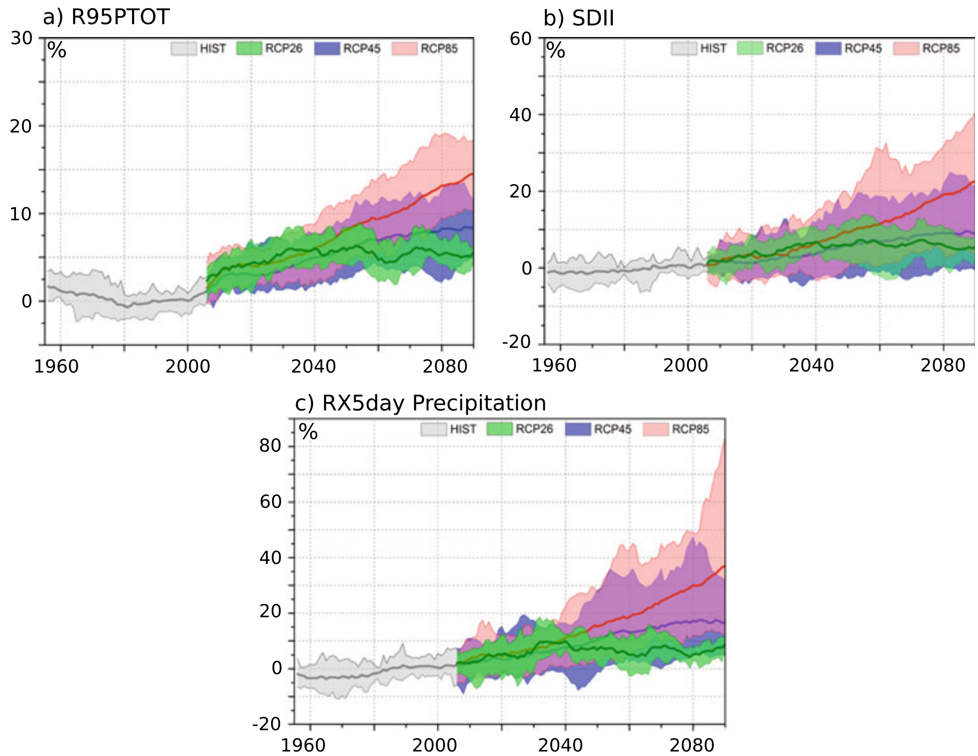

Fig. 3.12 Precipitation indices averaged over Indian land area **a** contribution of very wet days to total wet day precipitation (R95PTOT), **b** simple daily intensity index (SDII) and **c** maximum 5-day precipitation (RX5day) based on CORDEX South Asia multi-model ensemble. Changes are displayed relative to the reference period 1976–2005 (in %). Solid lines show the ensemble mean and the shading indicates the range among the individual RCMs. Time series are smoothed with a 20-year running mean

Change Detection and Indices (ETCCDI), computed with a consistent methodology for climate change simulations for different emission scenarios are discussed below.

Relative changes in the contribution of very wet days to total wet day precipitation (R95PTOT), the daily intensity index (SDII) and maximum 5-day precipitation (RX5day) with respect to 1976–2005 reference period are shown in Fig. 3.12. Similar to most of the global tropics, over Indian landmass as well, extreme precipitations are projected to increase throughout the twenty-first century. In RCP8.5, R95PTOT and SDII are projected to rise by 15 and 21%, by the end of the twenty-first century, whereas RX5day is projected to rise by 38%. The spread among ensemble members (shading in Fig. 3.12) is more in RCP8.5 comparing to RCP4.5 scenario all through the twenty-first century.

The spatial pattern of the projected multi-model ensemble means of the precipitation extremes identifies moderately higher increase in the contribution of very wet days to total wet day precipitation (R95PTOT; Fig. 3.13a), the daily intensity (SDII; Fig. 3.13b), and in the maximum 5-day precipitation (RX5day; Fig. 3.13c) are visible along the west coast, central and northern Indian states. Both the scenarios (RCP4.5 and RCP8.5) showed consistent results in the projected changes for both the near and far future. Even though the number of consecutive dry days (CDD; Fig. 3.13d) is increasing over various parts of India, the experiments provide a consensus only over the Indian peninsular region throughout the twenty-first century in the RCP8.5 scenario. The simultaneous increase in both CDD and RX5day indicates an increase of both dry and wet epochs along the west coast and the peninsular region of India. This analysis is in agreement with the study by Mukherjee et al. (2017), who noted the frequency of extreme precipitation from CMIP5-GCM shows an increasing trend prominent over southern India (Fig. 3.14). This result enhances our confidence in assessing a likelihood of an increase in future precipitation extremes over the Indian peninsula throughout the twenty-first century, under the propensity of global warming signals.

3.5 Uncertainties in Projected Precipitation Changes

Future climate projections are inherently saddled with uncertainties arising from multiple sources. These uncertainties are important to quantify in order to convey a realistic picture for future assessments, which are particularly useful at regional and sub-regional scales where local actions may form the basis for adaptation to expected changes in climate. Previous studies (Hawkins and Sutton 2009; Terray and Boe 2013) have identified three major sources of uncertainties in the future projections: (i) scenario uncertainty (ii) internal variability from chaotic nature of the climate system, and (iii) model related, i.e. how different climate models respond to the same forcing. Kirtman et al. (2013) showed that the uncertainty in near-term projections is mostly dominated by internal variability and model spread. This provides some of

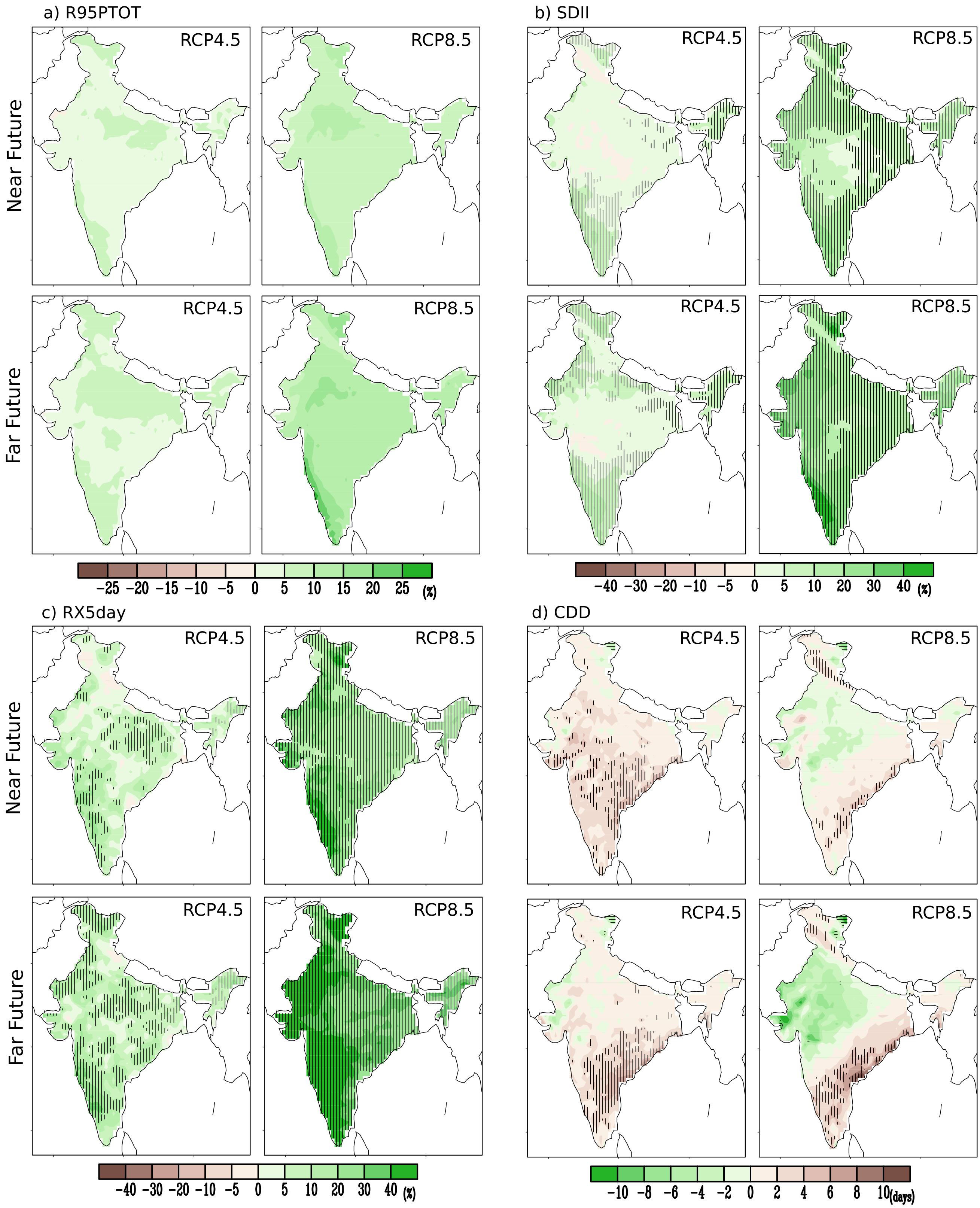

Fig. 3.13 Changes (%) in **a** contribution of very wet days to total wet day precipitation (R95PTOT) **b** simple daily intensity index (SDII), **c** maximum 5-day precipitation (RX5day), and **d** maximum number of consecutive dry days (CDD) for near future and far future from CORDEX South Asia multi-model ensemble mean with respect to the reference period 1976–2005. Striping indicates where at least 70% of the RCM realizations concur on the increase (vertical) or decrease (horizontal) in the future scenarios. The stapling is not plotted for R95PTOT as the RCM realizations have more than 70% consensus for increase

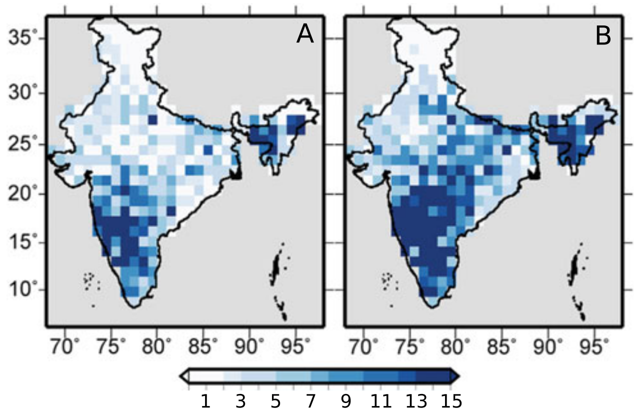

Fig. 3.14 Multi-model ensembles mean frequency of extreme precipitation events for the **a** Near Future and **b** Far future (Mukherjee et al. 2017)

the rationale for considering near and far future projections separately in most assessments.

Singh and AchutaRao (2018) quantified the uncertainties associated with models and internal variability over the Indian land region for each season using the RCP8.5 scenario of CMIP5 models. They showed that the uncertainty in precipitation change has a more complex picture such that uncertainty from internal variability persists and is quite large and comparable to model related uncertainty during the southwest and post-monsoon season by the end of the twenty-first century. The spatial heterogeneity in the uncertainty was comparatively more over the arid northwest region compared with the west-central region (part of the core monsoon area). This in a way enhances the confidence in the assessment of changes in precipitation over the central Indian region.

3.6 Knowledge Gaps

Though the modelling of the climate system has come a long way, there are still many issues that remain to be addressed. It is important to understand the complete monsoon system and to model the interactive processes that govern it. Identification of the coupled air-sea interactions, coupled land-atmosphere interactions, and flow-orography interactions that are critical in shaping the precipitation processes needs to be carried out. It is important to assess whether the model represents the phase transition in convection (shallow to deep to stratiform clouds). Bush et al. (2014) suggest that monsoon precipitation biases are sensitive to the entrainment and detrainment rates of convective parameterization. From observations and in models, it is important to identify the required thermodynamic conditions for convective phase transitions over the Asian monsoon region. There is lack of high-quality observations (atmosphere and ocean) over the monsoon-influenced regions to constrain the model physics. More realistic biogeophysical processes-based land surface models are needed to realistically assess the impact of land-use/land-cover change on the monsoon. The present models fail to adequately simulate the intra-seasonal variability of monsoon. There is limited knowledge about relative contributions of internal variability such as the Pacific Decadal Oscillation/Inter-decadal Pacific Oscillation (PDO/IPO) and external forcing in driving the historical evolution of monsoons and other modes of variability. It is challenging to project the behaviour of climate forcing. ENSO-Monsoon relationship is observed to have weakened in recent decades; however, models do not capture the ENSO behaviour in the future and how ENSO–monsoon relationship may evolve. Increased extremes and increased spatial variability have been observed in recent decades and also projected to increase, and it is likely due to regional forcings such as aerosols and land-use/urbanization changes.

3.7 Summary

Monsoons are the most important mode of seasonal climate variation in the tropics, and almost all parts of India receive more than 70% of the rains in summer monsoon season (June through September). The intensity, length and timing of monsoon are related to atmospheric moisture content, land–sea temperature contrast, land surface feedbacks, atmospheric aerosol loading and other factors. Overall, monsoonal rainfall is projected to become more intense in future, and to affect larger areas mainly due to increase in atmospheric moisture content with temperature. The temperature gradient between land and sea, regional distribution of land and ocean as well as topography play major role in monsoon.

Summer monsoon rainfall has decreased over India in the post-1950 period with more reduction in rainfall over the Indo-Gangetic plains and the Western Ghats. Global-scale anthropogenic forcings such as GHGs as well as regional-scale forcings such as aerosols and land-use/land-cover changes may have played a role in driving the changes observed in recent decades. The frequency of localized heavy rain occurrences has significantly increased over central India, which is partly attributed to changes in the availability of moisture due to greenhouse gas-based warming, aerosols, stability of the atmosphere and increasing urbanization. Extreme rains are concentrated around urban regions of India suggesting there is an urbanization feedback.

Global as well as regional models project an increase in seasonal mean rainfall over India while also projecting a weakening monsoon circulation. However, this weakening of circulation is compensated by increased atmospheric moisture content leading to more precipitation. Frequency of extreme precipitation events may increase all over India, and more prominently so over the central and southern parts as a response to enhanced warming. Monsoon onset dates are likely to be early or not to change much, and the monsoon retreat dates are likely to be delayed, resulting in lengthening of the monsoon season.

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