Monsoon low-pressure systems - the precipitation response to atmospheric warming

Silje Lund Sørland

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Acknowledgements

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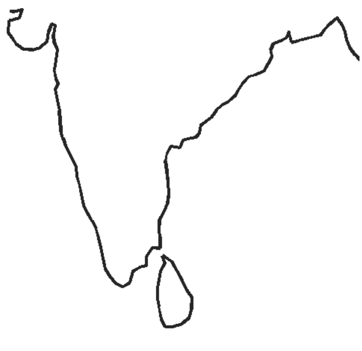

Abstract

The monsoon low-pressure systems (LPS) are amongst the most rain bearing synoptic scale systems that develop during the Indian monsoon, and they play a key role in generating extreme rainfall events in central India. This thesis consists of three scientific papers, which are devoted to the monsoon LPS, with focus on the precipitation associated with the systems. The main motivation for this thesis is to understand how the LPS precipitation may change in a warming world.

To examine this issue, a dataset of LPS developing during the Indian monsoon was generated, described in Paper I. A well know tracking algorithm was used to detect the LPS in the ERA-Interim reanalysis during the time period 1979-2010. The LPS that are connected to observed extreme rainfall events are selected, which resulted in a dataset of 39 LPS.

Next, high-resolution, convection-permitting, climate sensitivity simulations were performed on 10 cases chosen from the LPS dataset. Control runs are simulated with unperturbed ERA-Interim initial and lateral boundary conditions (LBC). Perturbed runs follow a surrogate climate change approach, in which a uniform temperature perturbation is specified to the LBC but the large-scale flow and relative humidity is unchanged. The difference between the control and perturbed simulations are therefore mainly due to the imposed warming and moisture changes as well as feedbacks to the synoptic scale flow.

The change in the mean precipitation following the LPS is described in Paper II, and Paper III focus on the change in the short-duration extreme precipitation released over central India and the runoff response. The results clearly show that in a warmer and more humid atmosphere, the LPS can produce more precipitation, precipitate with a higher precipitation rate and also bring precipitation further into the Indian continent. Based on these results we conclude that there may be an increased risk of more severe flood events in central India. The more than 2 x Clausius-Clapeyron scaling response in the precipitation is explained by the imposed specific humidity increase, a dynamic feedback giving stronger upward motions and a thermodynamic feedback decreasing the atmospheric stability. In the warmer runs the LPS are more intense with a higher propagation speed. The intensification of the storms seems to be interpretable in terms of the conditional instability of second kind mechanism: condensational heating increases along with low-level convergence and vertical velocity in response to temperature warming and moisture increases, and as a result the surface low deepens.

The results presented show that the precipitation associated with LPS are very sensitive to an increase in the atmospheric temperature and subsequent moisture increase. Changes in the atmospheric moisture content as a result of temperature warming are together with dynamical and thermodynamical feedbacks, affecting both the precipitation intensity and frequency.

List of papers

Paper I

Sørland, S. and A. Sorteberg (2015). The dynamic and thermodynamic structure of monsoon low-pressure systems during extreme rainfall events. *Submitted to TELLUS A.*

Paper II

Sørland, S., A. Sorteberg, C. Liu and R. Rasmussen (2015). The precipitation response in climate sensitivity experiments of monsoon low-pressure systems over India. *Submitted to Journal of Climate.*

Paper III

Sørland, S. and A. Sorteberg (2015). Low-pressure systems and extreme precipitation in central India: sensitivity to temperature change. *Submitted to Climate Dynamics.*

Contents

1 General introduction

With a population of 1.2 billion people today, India is expected to be the most populous country in the world by the mid 21st century (United Nations, 2012). This rapid population growth, together with a changing climate, is a threat to water security in the future (World Bank, 2013). The monsoon is the most dominant mode of atmospheric circulation over South Asia, giving wet summers and dry winters in India. Since up to 80% of the annual precipitation in India falls during the four monsoon months (June-September) (Ding and Sikka, 2006; Tyagi et al., 2012), it is evident that any changes in the intensity or frequency of the precipitation during the monsoon will affect agriculture, energy production and food supply (World Bank, 2013). One of the greatest potential risks to the people of India is therefore related to climate warming and the subsequent changes of the hydrological cycle, which may alter precipitation pattern, evapotranspiration and river flow, with potentially large impacts on the society.

The increase in atmospheric water vapor content with global warming is expected to change the hydrological cycle (e.g. Turner and Annamalai, 2012; IPCC, 2013). Possible increases in frequency and intensity of extreme rainfall events are thought to be one of the major consequences of changes in the hydrological cycle, and may yield a larger risk for severe flood events. The Indian subcontinent is particularly vulnerable to extreme rainfall events, since a large part of the population is poor, the infrastructure and crop yield is easily destructed by heavy precipitation events and early warning systems are not always reaching out to the affected people (World Bank, 2013).

Most of the extreme rainfall events in India occur during the monsoon, where monsoon low-pressure systems (LPS) play a key role in generating extreme rainfall events in the central Indian region (Goswami et al., 2006; Sikka, 2006; Pattanaik and Rajeevan, 2010). These synoptic scale systems, which are embedded in the large-scale monsoon flow, develop over India and the adjoining ocean and propagate in a northwestward direction towards the Indian continent (Saha and Sanders, 1981; Yoon and Chen, 2005; Sikka, 2006; Krishnamurthy and Ajayamohan, 2010). The LPS are categorized by the strength of the surface wind, where the most common ones are the weaker lows and more intense depressions (Tyagi et al., 2012). The LPS bring moisture into the Indian continent, which is released as precipitation along the propagation path. Recently, severe flood hazard have occurred in association with LPS, where the LPS brought high-moisture content into the region where the extreme rainfall occurred (e.g. the Pakistan flooding in 2010 (Houze et al., 2011); North-Indian flooding in 2012 (Joseph et al., 2014; Dube et al., 2014)).

The response of climate warming on the monsoon LPS remains an open research question. Several studies have reported a decrease in the frequency of the stronger monsoon depressions and an increase in the frequency of the weaker monsoon lows the last decades (Jadhav et al., 2009; Ajayamohan et al., 2010; Prajeesh et al., 2013). However, Cohen and Boos (2014) posed the question whether there have been a decreasing trend in the depressions, since they do not see a trend in the frequency of depressions in the distinctive LPS datasets obtained by using different automated methods. Studies suggest that storms in a warmer climate will be more intense with higher rainfall rates, because of the feedback latent heat may have on the storm (Trenberth et al., 2003; Westra et al., 2014). To know how sensitive the monsoon LPS are to the increase in the atmospheric temperature, would be of interest to the whole Indian community. The most important aspect is the sensitivity in the precipitation, since the Indian community is very dependent on the precipitation from the LPS, however, an increase in extreme rainfall events would have large socioeconomic consequences.

Since extremes are not well represented in global climate models, it is a challenge to study how precipitation extremes can change in a warming world. Downscaling global climate models with regional models is an important method when it comes to perform climate impact studies, and is invaluable for the end users (Maraun et al., 2010). However, several studies argue that since the climate models cannot reproduce the current climate accurately, the climate scenario projections are not reliable (e.g. Sabeerali et al., 2014; Sooraj et al., 2014), thus the downscaled projections will be influenced by biases in the climate models. Stowasser et al. (2009) downscaled a 4 times CO_2 scenario with a regional model over South Asia. They found that in the warmer climate, an increase in the number of flood days over central India could occur, as a result of more intense storms developing in the Bay of Bengal (Stowasser et al., 2009). In contrary, in another study the downscaled results suggested that there would be less precipitation in the central Indian region due to fewer depressions developing in the future scenario (Dobler and Ahrens, 2011). Whether the reduced number of LPS in the downscaled results reflects reality, or is due to the lack of ocean-atmosphere coupling in the regional model, is not know.

Objectives

In this thesis the LPS are investigated, where the motivation for the analyses is to understand how sensitive the LPS precipitation is to a warming world. First a dataset of monsoon LPS tracks that are associated with an observed extreme rainfall event over the Indian continent is generated. This is described in Paper I, where the relationship between LPS precipitation intensity and other meteorological parameters is investiated. Next, the sensitivity of LPS precipitation to temperature changes using a pseudo global warming approach on 10 LPS cases from the dataset generated in Paper I is simulated. The pseudo global warming method developed by Schär et al. (1996) consist of adding a temperature perturbation to the boundary conditions driving the regional model. The results are studied with respect to how the precipitation associated with the LPS changes in the warmer climate. It should be emphasized that the method prevents us from investigating the change in the frequency of the LPS, and instead the focus is on the change in the LPS precipitation. In Paper II the change in the mean precipitation associated with the monsoon LPS is studied in the warmer and more humid climate. Paper III focus on the short duration extreme precipitation that is released over the central India region, and the impact it has on the runoff. All simulations are conducted using high-resolution convection permitting simulations.

The main research questions are:

- How are the LPS that are connected to an observed extreme precipitation event represented in the ERA-Interim dataset?
- Which meteorological parameters are important for understanding the variability in the precipitation intensity associated with the monsoon LPS?

- How sensitive is the average precipitation associated with the LPS to an increase in the atmospheric temperature and moisture content?
- How sensitive is the extreme precipitation generated by LPS and associated runoff in the central India region to a warmer and more humid atmosphere?

The structure of this thesis is as follows: A theoretical background is given in section 2, where first the Indian monsoon is briefly described, followed by a more detailed description of the monsoon LPS. The theoretical background section is ended with a review of how the precipitation intensity can change with atmospheric warming. The data used in this thesis are presented in section 3, and the tools and methods are described in section 4. This thesis ends with a summary of the papers (section 5), followed by a discussion and future perspective in section 6.

2 Theoretical background

2.1 The Indian monsoon

Monsoon is a seasonal, atmospheric feature that covers large parts of the tropics, and around 65 % of the worlds population lives in a monsoon climate. The Indian monsoon is part of the South Asian monsoon system, and the monsoon onset over Kerala in southern India is normally in the beginning of June, and then the rainy season last throughout September (June-July-August-September: JJAS). Up to 70-80 % of the annual precipitation comes during these four months (Ding and Sikka, 2006; Tyagi et al., 2012), and the population of India is dependent on the monsoon rainfall for agriculture, industrial and hydropower purposes (Turner and Annamalai, 2012; World Bank, 2013). Figure 1 shows the monthly mean all-India rainfall, and it is clearly seen that most of the precipitation comes in JJAS.

The mean Indian monsoon

There are many ways to define the monsoon, but a traditional way has been the seasonal reversal of the wind direction (Gadgil, 2003). The Indian monsoon can be described as a large land-sea breeze, where the Coriolis force

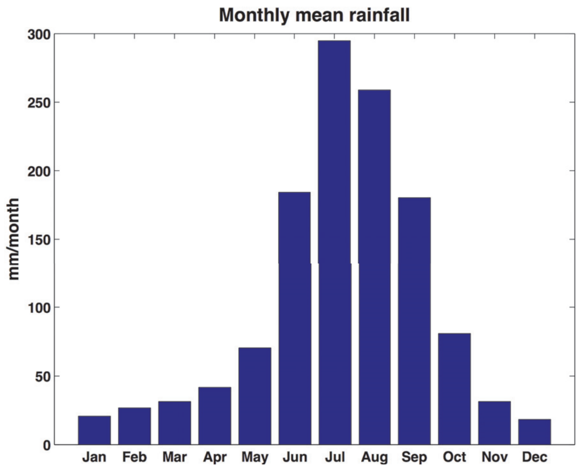

Figure 1: *All-India mean monthly rainfall distribution (1951-2010) from the IMD dataset (described in section 3).*

plays a role. Due to the annual variation in the incoming solar radiation and the different heat capacity of land and ocean, an inter-hemispherical circulation system arises. A thermal low develops over the northwestern Indian continent, giving rise to large cross-equatorial pressure gradients between the thermal low and the Mascarene High over the Indian Ocean in the southern hemisphere. This induces low-level winds that blows from the ocean in the winter hemisphere, towards the continent in the summer hemisphere. In the upper part the flow is reversed, blowing from the summer towards the winter hemisphere (e.g. Webster and Fasullo, 2003). The low-level and upper-level wind systems are shown in Figure 2, and the Coriolis force is the reason for why winds are deflected as seen in the figure. The low-level winds bring large amount of moisture from the warm tropical ocean, releasing as precipitation over South Asia. A comparison of the mean JJAS daily precipitation released over India from two different precipitation datasets (ERA-Interim and IMD, described in section 3) is shown in Figure 8, where it is seen large spatial variabilities in the precipitation over India. As a result of forced ascent over mountains and synoptic activity, the west coast, the east-northeast and central India are regions that receive more precipitation than compared to the rest of the country.

The circulation described above is how the Indian monsoon can be consid-

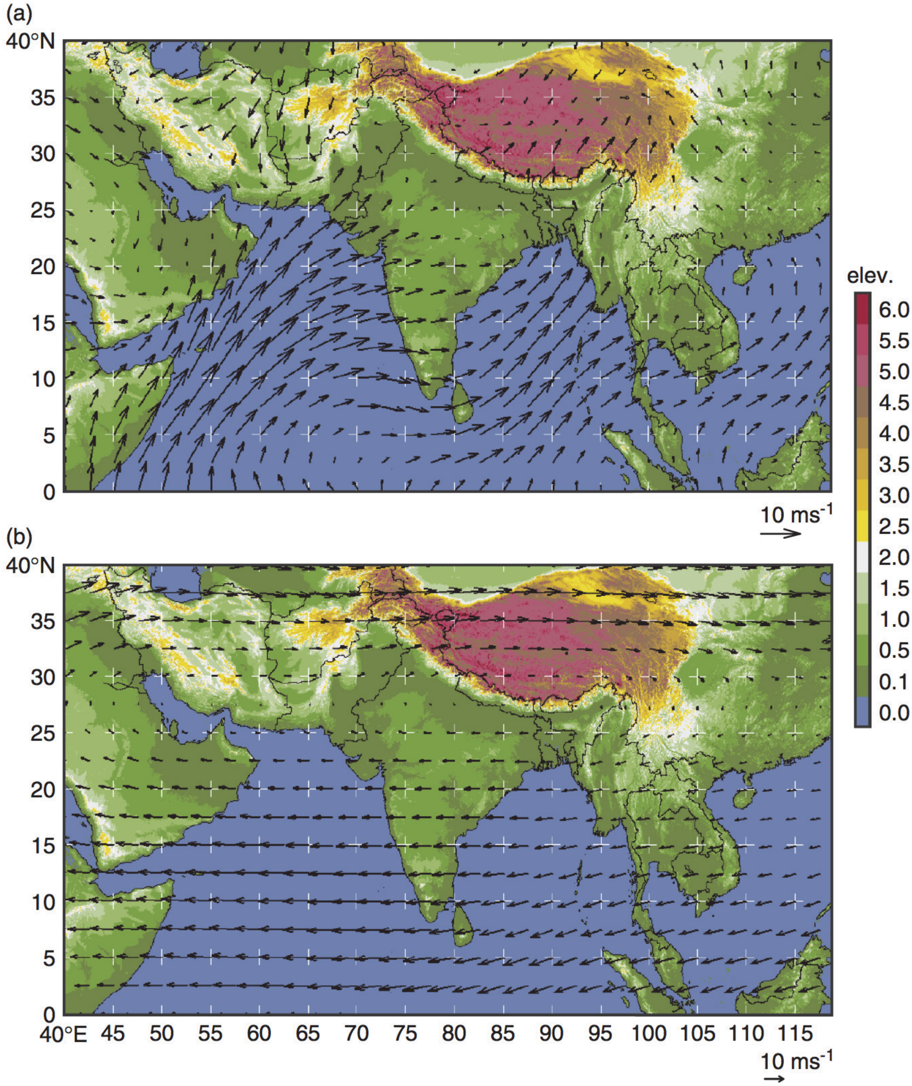

Figure 2: *Figure from Houze et al. (2007), where the mean wind during the summer monsoon at 1000hPa (a) at 200hPa (b) superimposed on topography (color shading [km]) is shown. The low-level winds in (a) is blowing from a high-pressure over the Indian Ocean in the southern hemisphere towards the thermal low over Northern India. The upperlevel winds is characterized by strong horizontal shear, with weak winds between the strong westerlies (at* $\sim 35°N$ *) and easterlies (at* $\sim 10°N$*).*

ered at the most basic level. However, there are several other mechanisms that also need to be considered to get the full picture of the monsoon system. The importance of the Himalayan and Tibetan Plateau has been studied in both the manner of the poleward extent of the monsoon precipitation, and how an elevated heating source over the continent can be important for the monsoon onset (Boos and Kuang, 2010; Turner and Annamalai, 2012). That the monsoon is part of the seasonal migration of the Intertropical Convergence Zone (ITCZ) is also a way of considering the Indian monsoon (Privé and Plumb, 2007). The monsoon trough is then seen as the northward migration of the Equatorial trough of the ITCZ (Tyagi et al., 2012).

Intraseasonal and inter-annual variations in the monsoon precipitation

The intraseasonal variability and spatial characteristics of the Indian monsoon rainfall can be explained by the intensification or displacement of the monsoon trough (Krishnamurthy and Shukla, 2000), and by the development of different synoptic scale systems (Ding and Sikka, 2006; Tyagi et al., 2012). The latitudinal oscillation of the monsoon trough has been related to the active-break cycle of the monsoon. The active period of the monsoon is when the trough lies over the northern part of India, bringing above normal precipitation to the central Indian region (Krishnamurthy and Shukla, 2000; Ding and Sikka, 2006). As the monsoon trough is shifted towards the foothills of the Himalayas, a break in the rainfall over (central) India is experienced. The temporal varation in the mean precipitation during one monsoon in the central Indian region is shown in Figure 3a, together with and a composite of the precipitation in India during active (Figure 3b) and break (Figure 3c) phases. The dynamics of the active-break cycle during the Indian monsoon is complex, where ocean - atmosphere interactions are suggested to be related to the movement of the convection (e.g. Joseph and Sabin, 2007).

The synoptic system such as mid-tropospheric cyclones, offshore troughs and monsoon low-pressure systems (LPS) are all contributing to distribute precipitation over the Indian continent. Mid-tropospheric cyclones have the maximum strength between 700 and 500hPa and are mostly located over the northeastern Arabian Sea and adjoining lands. The offshore troughs and vortices are found off the west coast of India (Ding and Sikka, 2006;

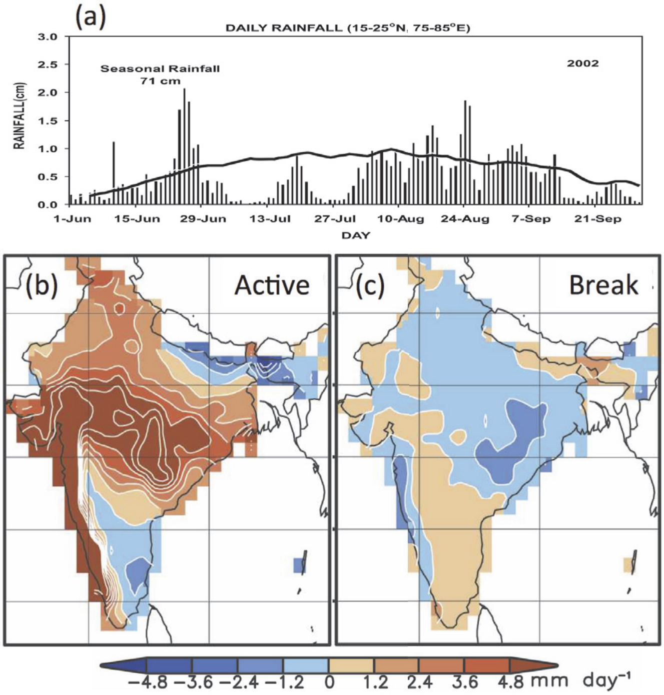

Figure 3: *The active-break cycle. (a) show the variation of the daily rainfall over central India (June - September 2002), where the monsoon was characterized by two long dry spells (breaks), from Rajeevan et al. (2010). Composite of daily rainfall anomalies (mm/day) for JJAS (1901-79) for the active phase (b) and 12 days before, which would correspond to break phases (c), from Krishnamurthy and Shukla (2007).*

Tyagi et al., 2012). The monsoon LPS is a collective name for low-pressure systems that mainly develop in the Bay of Bengal and propagate towards the Indian continent. The LPS are the main synoptic systems during the Indian monsoon, and are the key focus of this thesis, and will therefore be described in more detail in section 2.2.

The inter-annual variation of the Indian monsoon rainfall is because of changes in external forcings or internal variations. External forcing is a result of changes in the incoming solar radiation or in the atmospheric concentration of green house gasses or aerosols. Internal variations in the climate systems are driven by for instance changes in the snow cover over Eurasia or in variations in sea surface temperatures (SSTs) (Turner and Annamalai, 2012). The equatorial pacific SSTs are negatively correlated with the Indian monsoon rainfall, and this relation is referred to the ENSO-Monsoon interaction[1], and are considered to describe large part of the inter-annual variability of the Indian monsoon rainfall (e.g. Lau, 2003).

Has there been a change in the monsoon rainfall?

Even though there is clear evidence that the greenhouse-gas concentrations have increased since the pre-industrial time (IPCC, 2013), there is considerable uncertainty regarding the trends in the rainfall in Indian during the last century (Turner and Annamalai, 2012). Observations show large spatial (see Figure 8) and temporal variability in the rainfall over India (Guhathakurta and Rajeevan, 2008; Pattanaik and Rajeevan, 2010). Whereas some studies have noted that the all-India rainfall is quite robust and shows no significant trend over the last decades (e.g Goswami et al., 2006; Guhathakurta and Rajeevan, 2008), others have reported a declining trend in the all-India rainfall and parts of India since the 1950s (see the review by Turner and Annamalai, 2012). Turner and Annamalai (2012) emphasize that there are large uncertainties since the strength and sign of the trend depend on the region and period of study. Moreover, several studies suggest that there has been an increase in extreme rainfall events over India in recent decades (e.g. Sen Roy and Balling, 2004; Goswami et al., 2006; Rajeevan et al., 2008; Krishnamurthy et al., 2009). It should be stressed that the strength of these trends seems to be dependent on the definition chosen for an extreme rainfall event, and also there are some areas that experience an increase and some a decrease in extreme rainfall events. Figure 4a shows the spatial distribution of significant trends in extreme rainfall events over India, and Figure 4b shows the temporal variability of extreme rainfall in the central Indian region. An increasing trend of more heavy precipitation events in the central Indian region is see in Figure 4b, where a fixed treshold is used to pick out the extreme events. However, by using the 99th percentile as the treshold to pick out extreme rainfall events, there are spatial variabilities in the trend (Figure 4a). In central India there are areas where there is a positive trend,

[1] The SST anomalies in the equatorial pacific are related to the El Niño/Southern Oscillation (ENSO)

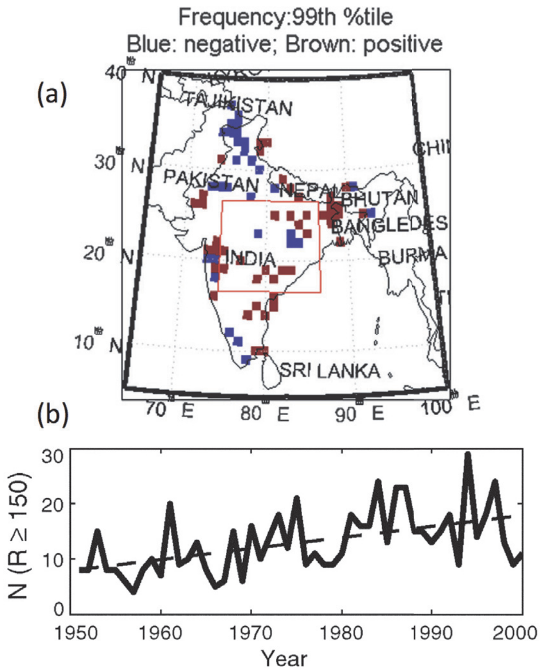

Figure 4: *Trend in extreme rainfall events. The spatial distribution of grids with a statistical significant trend, where blue indicate a decreasing trend and brown indicate a positive trend, for the frequency of exceedance of the 99th percentile (a). (b) shows the temporal variations of the number of events exceeding the daily rainfall intensity of 150mm ($R \geq 150$mm/day) in the central Indian region. (a) is from Krishnamurthy et al. (2009) and (b) is from Goswami et al. (2006), and the same time period and data is used in both studies. The red box shown in (a) correspond to the central Indian region considered in (b).*

some areas where no significant trend in the extreme rainfall events is found, and areas that have a negative trend. Thus the spread in the trends in rainfall events over India implies that trend analysis is something that should be done very carefully, and it is not easy to draw any conclusions.

Monsoon and climate change

The response of the increase in the greenhouse-gases on the monsoon system is complex. It is expected that the land surfaces will warm more rapidly than the ocean surfaces; therefore the land-sea contrast increases in most regions. However, the monsoon circulation is thought to weaken as the climate warms due to energy balance constraints in the tropical atmosphere (IPCC, 2013). The United Nations Intergovernmental Panel on Climate Change (IPCC) suggest that these changes in the atmospheric circulation can lead to regional changes in monsoon intensity, area and timing of the monsoon. The increase in the atmospheric temperature is associated with an increase in atmospheric moisture content, which can lead to an increase in total monsoon rainfall even though the strength of the monsoon circulation weakens or does not change. In addition, changes in the land-use and in atmospheric aerosols concentration is also thought to influence the monsoon system (IPCC, 2013).

2.2 Monsoon low-pressure systems

During the Indian monsoon, there are different weather systems that contribute to the total monsoon rainfall. The synoptic scale monsoon low-pressure systems (LPS) are one of the most efficient rain-producing systems developing, and they have been found to produce about half of the Indian monsoon rainfall (Yoon and Chen, 2005). The Indian Meteorological Department (IMD) categorize the LPS based on the surface winds, where the wind strengths in the ranges of up to < 8.5, 8.5-16.5, 17-23.5, 24-31.5 and greater that 32 m/s correspond to systems as lows, depressions, cyclonic storms, severe cyclonic storms and hurricanes (Tyagi et al., 2012). During the monsoon, the LPS do not develop into hurricanes, due to the strong vertical wind shear in the background flow (Tyagi et al., 2012). The thermal structure of the monsoon LPS is characterized by a cold anomaly at lower levels and warm core aloft. Evaporative cooling from the falling precipitation is suggested to generate the low-level cold core. The monsoon LPS have a typical horizontal extent of 1000-2000 km radius, and a well-defined life cycle of 4-7 days. The systems develop in the Bay of Bengal (BoB), the Arabian Sea or over land and normally propagate in a northwestward direction towards the Indian continent (Saha and Sanders, 1981; Yoon and Chen, 2005; Sikka, 2006; Krishnamurthy and Ajayamohan, 2010). Figure 5 shows the tracks of the LPS developed during the monsoon in the period 1888-2003. Most of the systems develop in Bay of Bengal, whereas some form over Arabian Sea and land.

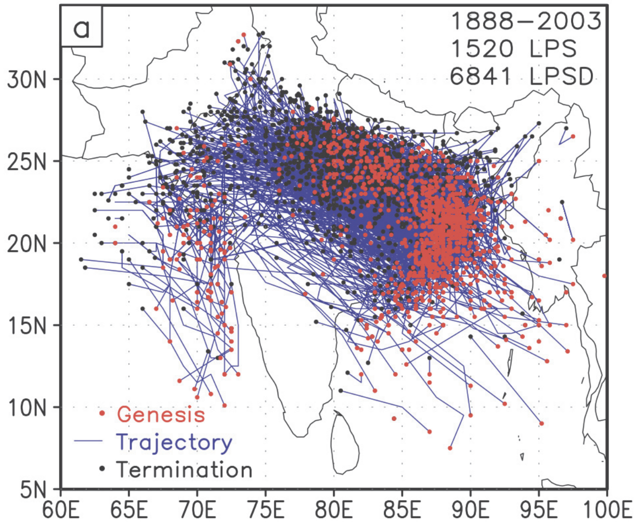

Figure 5: *The trajectories (blue line) of LPS formed during the monsoon (JJAS) from 1888-2003. The genesis is the red dot, while the termination is the black dot (Krishnamurthy and Ajayamohan, 2010).*

The evolution of the life cycle for a typical monsoon LPS is shown in Figure 6. It is seen that the low-pressure develops in the BoB and propagates in a northwestward direction towards the Indian continent. From the 18th -19th June the systems covers large parts of central India, with convective cells around the center of the low, mainly located in the southwest sector relative to the low-pressure center. From Figure 6 it is clearly seen that the low-pressure propagate deeply into the Indian continent, and is in this way contributing to distribute precipitation over large parts of India. As there are on average 13-14 LPS developing each monsoon season (Sikka, 2006), it is evident that they are important contributors of the total precipitation during the monsoon.

There are more LPS developing in the active periods than the break periods of the monsoon (Goswami et al., 2003; Krishnamurthy and Shukla, 2007). The paths of the LPS during the active period are clustered along the monsoon trough, therefore the central Indian region receives more pre-

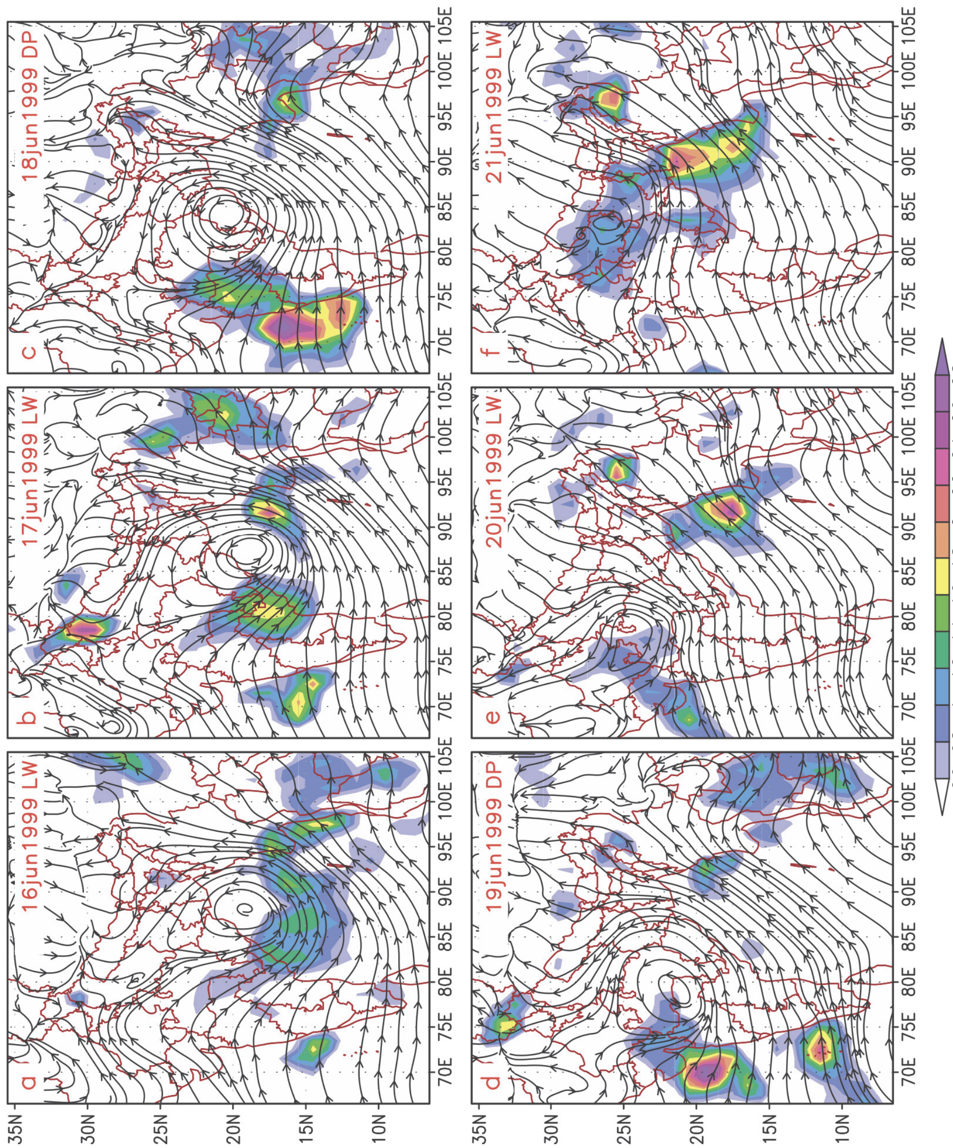

Figure 6: *The life evolution of a typical LPS, taken from Krishnamurthy and Ajayamohan (2010). The black lines are the 850hPa streamlines [m/s], and the shading is the precipitation [mm/h]. The intensity of the LPS is given at the top of each subfigure (LW: low, DP: depression).*

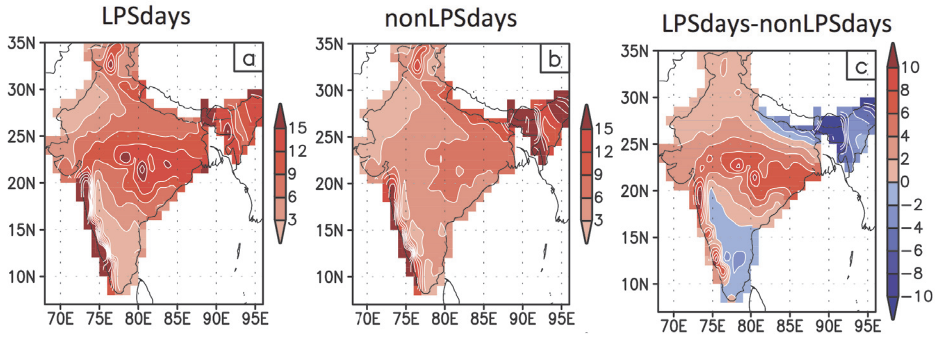

Figure 7: *Composite of the daily precipitation [mm/day] for (a) days when LPS are present, during non-LPS days (b), and the difference between the composite of LPS days and non-LPS days (c), taken from Krishnamurthy and Ajayamohan (2010). Note the similarity between (c) and Figure 3c, which show the composite of the precipitation anomaly during active phases.*

cipitation from these systems than compared to the rest of the continent (Goswami et al., 2003). By performing a composite study of 1520 LPS that developed from 1888-2003 (see Figure 5), Krishnamurthy and Ajayamohan (2010) showed that during days when the LPS are present (LPS days), central, southwest and northern part of India receive up to 10 mm more precipitation each day, compared to the rest of the Indian subcontinent. Meanwhile, during days when there are no LPS present, southeast and northeast India receive approximately 6-8 mm more precipitation each day (Krishnamurthy and Ajayamohan, 2010), shown in Figure 7.

Multiple studies have reported that the frequency of the weaker LPS (lows) during the Indian monsoon have shown an increasing trend the last decades, while the stronger monsoon LPS (depressions) have experienced a declining trend (Jadhav et al., 2009; Ajayamohan et al., 2010; Prajeesh et al., 2013). It is not obvious why this shift in the intensity of the monsoon LPS is observed, but several mechanisms which are related to climate change and increase in aerosol consentration, are suggested: (1) The increase in the SST in the Bay of Bengal (BoB) is not favorable to intensify the weaker lows into depressions, since the convection is being diminished when the SST is above $29.0°C$ (Jadhav et al., 2009); (2) A reduction of the mid-tropospheric humidity is reducing the frequency of the monsoon depressions (Prajeesh et al., 2013); (3) The increase in atmospheric aerosol from air pollution acts to destroy the organization of the convection around the LPS, which weakens the depression, or inhibits the low from developing into a depression (Krishnamurti et al., 2013). However, Cohen and Boos (2014) posed the question whether there have been a decreasing trend in the depressions, since they do not see a trend in the frequency of depressions in the distinctive LPS datasets constructed by using different automated detection methods.

The monsoon LPS are related to extreme rainfall events in the central Indian region (Goswami et al., 2006; Sikka, 2006; Pattanaik and Rajeevan, 2010). These extreme rainfall events can cause damages to life and property through flooding and landslides. An increase in the extreme rainfall events in the central Indian region the last decades is suggested (Goswami et al., 2006; Pattanaik and Rajeevan, 2010), and this increase has occurred despite the decrease in the frequency of the stronger monsoon depression implied by different studies (Jadhav et al., 2009; Ajayamohan et al., 2010; Prajeesh et al., 2013). The IPCC report relate the increase in the atmospheric moisture content or the warmer SSTs in the tropical Indian Ocean (Rajeevan et al., 2008) to the increase in the extreme rainfall events in the central Indian region (IPCC, 2013). However, as discussed in section 2.1 and above, the increasing trend in the extreme rainfall events and the decreasing trend in the monsoon depression comes with large uncertainties.

The LPS propogate towards the west-northwest, against the mean low-level westerly flow. This westward propagation is connected to the low-level convergence, which is mainly in the west-northwestward sector relative to the low-pressure center (Godbole, 1977). The reason for the low-level convergence and the corresponding strong updraft ahead of the low-pressure center can be due to boundary layer friction, warm air advection, vorticity advection and cumulus processes (Sikka, 2006; Tyagi et al., 2012). There have been several studies where the dynamics of the monsoon LPS and the westward propagation have been studied (Sanders, 1984; Chen et al., 2005; Boos et al., 2014). By using quasi-geostrophic theory it has been suggested how the upward branch west of the depression center is maintained through latent heat release, generating vortex stretching, allowing the depression to propagate westward (Sanders, 1984; Chen et al., 2005). However, recently, Boos et al. (2014) proposed a new view of the LPS dynamics, where they argue that the monsoon depressions propagate westward by nonlinear, horizontal adiabatic advection. Since air has to conserve its angular momentum, any changes in the coriolis force in the air surrounding the cyclone, will generate local cyclonic and anticyclonic gyres, letting the cyclone drift poleward and westward (called the beta drift). Boos et al. (2014) suggest the colocation of the precipitation and the upward ascent only has a minor contribution on the dynamics of the LSP. This new hypothesis just emphasizes that the understanding of the LPS dynamics and thermodynamics is not complete, and there is room for more research to understand this complex dynamics, which can be influenced by processes occurring on different scales.

To the west of the LPS, there is warm air advection from the north, and to the east of the cyclone center there is cold-air advection from the south (Saha and Chang, 1983), which suggest that baroclinic processes are important for the LPS dynamics. By analyzing the thermal budget of a monsoon depression, Saha and Saha (1988) propose that even though the temperature advection is small, it could be contributing to the initiation of vertical motion. Shukla (1978) highlighted the importance of the combination of the barotropic-baroclinic instability and the Conditional-Instability of Second Kind (CISK) for the development of the depressions, and suggested the primary driving mechanism for the growth of the monsoon depressions to be the CISK mechanism. The CISK-concept was first used to describe the growth of tropical hurricanes (Charney and Eliassen, 1964; Lin, 2007), and the theory portrays how a tropical disturbance can grow in a conditional unstable atmosphere through cooperation between the cumulus convection and the large-scale circulation. When there is low-level synoptic convergence and associated boundary layer friction, air with high moisture content is pumped upward in the atmospheric column (Ekman pumping). When the air reaches the cloud base (i.e. saturation), and eventually the level of free convection, latent heat will be released in the atmospheric column. This cumulus convection process will enhance the low-level convergence, which leads to more moisture pumped upward in the atmospheric column, and in this way is a self-excising process, until the vertical velocities (and low level convergences) are suppressed or there is a reduction in the moisture.

2.3 Precipitation and climate change

One of the most important variables within meteorology and hydrology is precipitation. To understand precipitation processes and predict precipitation accurately is extremely important, because of the huge impact it can have on the society. For instance, associated with severe storms there are large rainfall rates, causing flood events, which affect livelihood and lead to large economic losses. On the contrary, a depletion of precipitation over time causes increased frequency of droughts and a reduction in the water supply, which also has crucial impact on the society.

To predict how precipitation changes with a warming climate is a challenge because the precipitation processes are nonlinearly associated through dynamic, thermodynamic, cloud microphysical and radiative processes (Li and Gao, 2012). As an air parcel rises and expands, the air parcel will be cooled adiabatically. When the temperature of the air parcel is reduced to the temperature at saturation, the water vapor will condense into water droplets, produce clouds and eventually precipitation (e.g. Wallace and Hobbs, 2006). Clearly the vertical velocity is one of the main reasons for generating precipitation, however because of the latent heat release during phase changes of water, and the evaporative cooling from the falling rain, the precipitation is interlinked through atmospheric dynamics and thermodynamics processes, and these complex interactions makes is difficult to assess how the precipitation can change with global warming. A change in the frequency of the precipitation is often dependent on the atmospheric circulation and the moisture convergences generated, whereas the intensity of the precipitation is closer connected to the atmospheric moisture content that will be further complicated by cloud microphysical and radiative processes that will influence the timing and/or amount of the precipitation generated.

Scaling of precipitation

Based on thermodynamic properties, expressions of the precipitation intensity have been derived. A common assumption of the change in the precipitation intensity with temperature is related to the water holding capacity of the atmosphere, which is given by the Clausius-Clapeyron (CC) equation:

$$\frac{dq_s}{q_s} = \frac{L_v dT}{R_v T^2} \tag{1}$$

Here the relation between the saturation vapor pressure (e_s) and saturation specific humidity (q_s) is used, where $q_s = 0.622\frac{e_s}{p}$, p is the atmospheric pressure, L_v is the latent heat of vaporization, R_v is the gas constant and T is the temperature. The fractional increase in the saturation specific humidity is in the range of 6-7 % when the temperature is increasing with 1 K (i.e. 6-7%/K). Thus if the relative humidity remains constant under climate change (IPCC, 2013), the change in the specific humidity will follow the CC-relation, and so will the precipitation, if it should change as the atmospheric column integrated water vapor content. This precipitation change, which is constrained by the moisture availability, is referred to as the CC-scaling.

By using climate models different studies have shown that the global mean precipitation does not scale as the CC-relation, but at a lower rate, around 3.4 %/K (e.g. Allen and Ingram, 2002). This is because latent heat is released as water vapor condenses. Thus the change in the global mean precipitation is constrained by the energy budget, and not by availability of the moisture content. O'Gorman and Schneider (2009) show this by deriving a scaling of the condensation rate, c (i.e. the condensate rate: $c = \frac{dq_s}{dt}$ when $\frac{dq_s}{dt} < 0$). Assuming that the precipitation is analog with the condensation rate, which is given by:

$$c \sim \omega \frac{dq_s}{dp}\bigg|_{\theta_*} \tag{2}$$

ω is the vertical velocity in pressure coordinate, p is the pressure, and the moist-adiabatic derivative of saturation specific humidity ($\frac{dq_s}{dp}$) must be taken along a constant equivalent potential temperature (θ^*) to allow for the warming effect of latent heat release (O'Gorman and Schneider, 2009). Thus, the condensation rate given above will not have a temperature scaling as the CC-scaling because the latent heat released at higher temperature will moderate the condensation rate, given that the relative humidity and vertical velocity will not change.

This last scaling approach is appropriate to describe how the global mean precipitation intensity can change with warming. Though, it is important to distinguish between changes in global and local precipitation changes. Romps (2011) nicely presents how the local and global precipitation intensities change in high-resolution simulations during doubling of CO_2 concentrations. He find the global mean precipitation intensity to increase by 3%/K, and the local precipitation to increase by 6%/K. Even though the local precipitation change is in the order of the CC-scaling, this is more by chance than the explanation. The contribution to the local precipitation change is partly from increases in the vertical velocity, the vertical water vapor gradient and the cloud mass (Romps, 2011). Romps (2011) also finds the most intense precipitation to increase by 7-8%/K, and these events are generated by convective plumes that are faster and wider, but they are developing less frequently in the warmer simulations.

Analog to the results from Romps (2011), Trenberth et al. (2003) hypothesize that in a warmer climate there will be fewer storms, but the storms will be more intense, most likely due to the increase in the latent heat release. Associated with the storms there will be larger precipitation rates, however a reduction in the lifetime can be expected (Trenberth et al., 2003). This effect on storms in a warmer climate is conceptualized in Westra et al. (2014) by explaining it through a cloud feedback mechanism. When the only difference of the storm with warming is the moisture content, the precipitation intensity will change as the CC-scaling. According to Westra et al. (2014), if the storm is becoming more vigorous with stronger updrafts, but the horizontal extent remains the same, the lifetime of the storm will be reduced. The precipitation intensity during the storm will be larger than the CC-scaling, but since the storm has a reduced lifetime, the total accumulated precipitation will scale as the CC-relation. However, if the area of the storm is increasing together with a stronger updraft, even more moisture can be fed into the storm. For this situation, it is not clear if the lifetime of the storm will be reduced, but nevertheless, the precipitation intensity during the storm and the total accumulated precipitation will scale beyond the CC-scaling (Westra et al., 2014).

With global warming the extreme precipitation is expected to increase at a larger rate than the mean precipitation (O'Gorman and Schneider, 2009; Pall et al., 2007). Extreme precipitation events are assumed to be constrained by the moisture availability, therefore it is suggested that the extreme precipitation will scale as the CC-relation, or even at a greater rate if the circulation also increases (Allen and Ingram, 2002; Trenberth et al., 2003). In a review by Westra et al. (2014), the globally averaged daily extreme precipitation during the 20th and early 21th century is reported to have increased by 5.9-7.7%/K, and it is also found that the extreme precipitation on sub-daily time scale has been intensified significantly. Loriaux et al. (2013) shows that the stratiform extreme precipitation, which has a longer time-duration and is more common in colder weather, is following the CC-scaling, while convective extreme precipitation that has a shorter time duration and appears in warmer weather, has a fractional increase of 2 times the CC-ration. Thus, it could be expected that the tropical extreme precipitation, which are predominaly of convective charater, should increase at a super-CC scaling (O'Gorman, 2012).

Diagnosing the precipitation

To evaluate how the precipitation can change with global warming, let us take one step back, and start by considering which parameters are important for the precipitation intensity. By assuming saturation and neglecting the microphysical delay, the precipitation intensity is dependent on the change in condensed water with time, integrated over the vertical extent of the atmosphere where the condensation occur (cloud bottom to cloud top). The precipitation intensity is then approximated as:

$$P \approx \begin{cases} -\displaystyle\int_{P_{C_b}}^{P_{C_t}} \frac{dq_s}{dt}\,dp & \text{if } \frac{dq_s}{dt} < 0, \\ \\ 0 & \text{if } \frac{dq_s}{dt} \geq 0. \end{cases} \tag{3}$$

q_s is the saturated specific humidity, p is the pressure, P_{C_t} and P_{C_b} is the pressure at the top and bottom of the cloud. The precipitation in the above expression is given as the cloud integrated amount of condensed water, where the precipitation efficiency is assumed to be unity, thus the condensed water falls out instantaneously without being transported or evaporated before it reaches the ground.

The saturated specific humidity is a function of the temperature through the CC- equation ($\frac{dq_s}{dT}$), the temperature is a function of the pressure through the atmospheric stability ($\frac{dT}{dp}$) and the pressure is a function of time, given by the vertical velocity in pressure coordinate ($\frac{dp}{dt}$). Thus by using the chain rule, the condensed water (and the precipitation) can be given by:

$$c = \frac{dq_s}{dt} = \frac{dq_s}{dT}\frac{dT}{dp}\frac{dp}{dt} \text{ when } \frac{dq_s}{dt} < 0. \tag{4}$$

If the atmospheric stability and vertical velocity is assumed to be constant with global warming, the change in the condensed water will only be dependent on the CC-relation, thus the amount of condensed water will increase with 6-7%/K. By including changes in the atmospheric stability as a result of for instance latent heat release, the change in the condensed water with global warming is the same as shown by Allen and Ingram (2002) and O'Gorman and Schneider (2009). However, if the vertical velocity is not constant with climate changes, this scaling assumption may not hold. Thus to assess the change in the condensation with warming, changes in the saturated specific humidity, atmospheric stability and vertical velocity has to be included.

Saturated specific humidity is directly related to the atmospheric temperature through the CC-equation. During rainfall events, the atmospheric temperature can change as a result of for instance evaporative cooling from the falling precipitation, and this will influence the scaling result (e.g. Westra et al., 2014). There have been studies that have used alternative temperatures as for example the dew point temperature (e.g. Loriaux et al., 2013). The dew point temperature represents the absolute specific humidity of the atmosphere, thus it could be suggested to use the actual moisture content of the atmosphere rather than the temperature (or the saturated specific humidity) when predicting the precipitation change (Westra et al., 2014).

3 Data

Retrospective-analysis (reanalysis) are systems that merge weather observations and models into one global 4-dimensional dataset. By using numerical models the weather observations are assimilated into global gridded values. Reanalysis dataset have been extensively used in several climate studies, and the users benefit greatly from the long time series of the global coverage of data. However, not all the fields that are produced come with the same reliability. The variables that are directly assimilated from observations gives more reliable estimate than the variables that are derived from the model. Variables describing the large-scale atmospheric circulation (e.g. temperature) are directly assimilated from observation data, whereas precipitation is a diagnosed variable. Precipitation is therefore not only dependent on the quality of the observation assimilated into the system, but the model physics will also introduce uncertainties. Precipitation from reanalysis have shown a poorer skill in the tropics than in the extra-tropics, and this can be related to how the convection is parameterized, since convection is more dominant in the tropics than at higher latitudes (e.g. Bosilovich et al., 2008).

This thesis uses a reanalysis from the European Centre for Medium Range Weather Forecast (ECMWF); ERA-interim. The ERA-Interim is an atmospheric reanalysis product covering the period from 1 January 1979 and onwards (Dee et al., 2011). The dataset is given on a spectral resolution of T255, which is approximately 80 km, on 60 vertical levels, and is produced with a 4-dimentional variational assimilation system (4D-Var). ERA-Interim is used as the lateral boundary conditions for the WRF simulations analyzed in Paper II and III. In Paper I ERA-Interim is used to detect the monsoon LPS, and it is also performed analysis on several different parameters from the reanalysis dataset, including the diagnosed precipitation.

Lin et al. (2014) analyzed 5 different global reanalysis datasets and assessed their performance in reproducing the climatology, inter-annual variation and long-term trend of global monsoon precipitation. They found that the ERA-Interim is the reanalysis dataset that shows the highest skill in reproducing the climatology and the inter-annual variability of monsoon precipitation. They suggest that the better performance of the ERA-Interim precipitation is a result of the advanced assimilation method. However, Lin et al. (2014) emphasizes that ERA-Interim overestimates the moderate rain substantially. This overestimation of the moderate rain is also found in our investigations (Paper I). That the ERA-Interim is underestimating the intensity of extreme precipitation is also found, which is confirmed by previous studies (Pfahl and Wernli, 2012). The mean JJAS and 99.5th percentile of the precipitation from the ERA-Interim is shown in Figure 8, where it is compared with the observational based Indian Meteorological Department rainfall, introduced in the next subsection. The underestimation of the extreme precipitation is evident when comparing the two datasets.

Due to the large uncertainty in the quality of the precipitation, two other precipitation dataset are included: the Indian Meteorological Department rainfall (IMD) data (Rajeevan et al., 2006), and the Tropical Rainfall Measuring Mission precipitation (Huffman et al., 2007). The IMD is a rain gauge based daily gridded dataset on 1 ° x 1 ° covering the Indian region, available from 1951-2010. IMD is used in Paper I and II, where in Paper I the IMD-rainfall is used as the "truth" to pick out the extreme rainfall events that are found in both the IMD rainfall and the ERA-Interim precipitation. In Paper

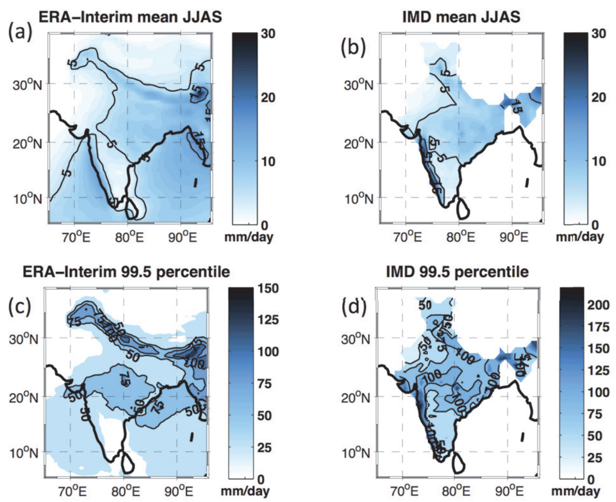

Figure 8: *The mean daily JJAS precipitation from ERA-Interim (a) and IMD (b), and the 99.5th percentile of the mean JJAS daily precipitation from ERA-Interim (c) and IMD (d), for 1979-2010. Note the different color bars in (c) and (d). In (a) and (b) [(c) and (d)] the 5 and 15 [50, 75 and 100] mm/day contours are shown in black.*

II the WRF precipitation is evaluated with the IMD rainfall in the central Indian region. TRMM is a satellite monitoring system with the purpose to monitore and study tropical rainfall (Huffman et al., 2007). It has a spatial resolution of 0.25° x 0.25 ° with a 3-hour time resolution, covering the tropics and part of the extra-tropics (50°N-50°S). TRMM is only used in Paper I, where is is compared with the ERA-Interim precipitation associated with the monsoon LPS.

4 Tools and methods

4.1 Tracking monsoon LPS

Detection of a specific weather system in model outputs/reanalysis by using an automated method and objective criteria has been used in several studies (i.e. Yoon and Chen, 2005; Bengtsson et al., 2007; Stowasser et al., 2009; Pfahl and Wernli, 2012). This thesis uses the TRACK program developed by Hodges (1994, 1995, 1999), which is a Lagrangian feature-tracking algorithm. The basis of the method is to search for maxima or minima in a meteorological field (e.g. relative vorticity, geopotential height or MSLP), and the feature points that exceed a prescribed threshold are tracked and connected to a trajectory by minimizing a cost function (Hodges, 1994, 1995, 1999). The choice of the meteorological field should be done on the basis of what scale is to be detected (i.e. tracked). For instance, the MSLP is a better field to track the larger scales, since it is distinctively influenced by stronger background flow, where large spatial scales and relative slower moving systems dominate. The background flow is less dominate in the relative vorticity field, and is therefore a better choice to use when identifying smaller scale systems.

The dataset of the monsoon LPS used in this thesis is generated using the 850hPa relative vorticity field truncated to the spectral resolution of T42 with a time resolution of 6 hours. The vorticity threshold is $0.5x10^{-5}s^{-1}$, and only the trajectories that travel minimum 5° and last for more than 48 hours are retained. The removal of the background flow has been a common method to filter out features smaller or larger than on the scale of the weather phenomena to be detected. However, here the background flow in the relative vorticity field is not removed because the aim was to detect the features initially, and remove the tracking errors by introducing constraints based on characteristics of the monsoon LPS. These criteria mentioned here are called initial constraints. These constraints are typical to use when performing an automated tracking method, where the only difference is the magnitude of the vorticity threshold, lifetime and minimum displacement. The next constraints we call adaptive constraints, which are criteria chosen based on the characteristics of the monsoon LPS and the area and season we are focusing on. General details about the method for detecting the monsoon LPS is described fully in Paper I, and a summary of all the criteria used is listed in Table I. Applying the algorithm on the whole time period of study (1979-2010), and using all the criteria described above leave us with 133 systems during the 32 years of study, with 4.2 trajectories detected each season on average.

To evaluate the results from the automated TRACK method and tune the different criteria, we compared our trajectories with trajectories in the Sikka dataset (Sikka, 2006). The Sikka dataset consists of statistics of the position and time duration of monsoon LPS that developed over India and the adjoining oceans from 1984-2003, and is constructed by analyzing synoptic weather charts of the MSLP and the surface wind speed. Based on these two parameters the monsoon LPS have been classified into a low, depression, deep depression, cyclonic storm or severe storm.

During the 20 years studied by Sikka (2006) 274 LPS are found (on average 13.7 LPS each season), where 45 are categorized as depressions or deep depressions (2.25 each season). The coincidence of the trajectories found by our method, and with the observed trajectories found in Sikka (2006) is very good, however, we realized we were left with not only depressions, but also the weaker systems monsoon lows. Thus, based on these results it cannot be concluded that the method and associated constraints used here is capable of detecting the climatology of the monsoon depressions, but it generate a dataset of monsoon LPS that have the same thermal structure. The method used in Sikka (2006) to pick out the LPS, does not consider the thermal structure of the low. The number of LPS found by our method and the number of LPS found in Sikka (2006) should therefore not necessarily

		Criteria
Initial constraints	Vorticity treshold	The feature points in the 850 hPa T42 vorticity need to exceed a threshold of 0.5 x10-5 s-1.
	Displacement treshold	A minimum displacement of $5°$ during its lifetime.
	Lifetime	The detected feature needs to have a lifetime of more than 48 h.
Adaptive constraints	MSLP treshold	A MSLP minimum search around each feature point within a radius of $5°$ is performed. The features that have a true minimum for at least four consecutive time steps are picked out.
	Cold/warm core	Identification of the systems that are cold core at lower levels and warm core aloft, where the T63 850 hPa relative vorticity maximum needs to exceed a threshold of 1 x10-5 s-1. To check for the cold and warm core, the vorticity difference in the lower layer (between 850 – 500 hPa) is required to be positive, while the vorticity difference between the upper layer (500 – 250 hPa) must be negative. This must be true for one time step.
	Topography	All the trajectories that developed above 700 m are removed, which may be a result extrapolated values in the reanalysis.
	Direction of movement	Picking out the systems that move in a northwestward direction.
	Monsoon season	The summer months (JJAS) are studied.
	Indian region	Area of study: latitude $10 - 30° N$ and longitude $65 - 100° E$.

be the same. To compare a climatology constructed with different methods from different datasets is not trivial, and this is clearly shown by Cohen and Boos (2014). They compare the number of depressions detected in different dataset, and the annual frequency and variability differs depending on the detection method and dataset used. Based on their results, they pose the question whether there have been a decreasing trend in the depression the last decades, since they do not see a trend in the frequency of depressions in the different LPS datasets obtained by using different automated methods.

4.2 Dynamical downscaling by using the PGW approach

Instead of performing the traditional method of dynamical downscaling on projections from climate models, the Pseudo Global Warming approach (PGW) is used in this thesis. By adopting a realized or simulated atmospheric flow evolution (i.e. reanalysis data) as LBC, and add a climate change signal to the driving boundary conditions, the results from the regional model are investigated in a future climate scenario. This PGW approach is appropriate for sensitivity studies, and it provides a first order estimate of the impact from climate change. The approach introduced by Schär et al. (1996) is used, called the surrogate climate-change simulation. This method separates the effect of the increase in atmospheric moisture from changes in the atmospheric circulation, so the results can be studied with respect to how changes in the thermodynamic fields affect the precipitation.

The PGW approach is used on 10 different monsoon LPS cases, chosen from the dataset generated in Paper I. In Paper II and III the results from the surrogate climate change simulations is presented. Here the methodology behind the sensitivity experiments is presented. ERA-Interim is used as initial and latera boundary conditions for the simulations. First we perform a control simulation without any perturbations added to the boundary conditions (the control run). Next, a temperature perturbation, ΔT, is added to the temperature fields on all pressure levels in the atmosphere, the SST, the skin-temperature and the soil temperature fields in the ERA-Interim initial and lateral boundary conditions. The new temperature fields are then given as:

$$T_{\Delta T} = T_{CTR} + \Delta T, \tag{5}$$

where ΔT is the temperature pertubation. We perfrom two perturbed runs, with a temperature pertubation of $\Delta T = 2K$ and $\Delta T = 4K$, referred to the +2K and +4K runs, respectively. It shold be stressed that we are perturbing the temperature and not the virtual temperature as in Schär et al. (1996). This is a valid approximation, since the temperature change due to the change in the humidity, is very small compared to the total temperature increase.

The relative humidity (RH) is given as the fraction between the specific humidity (q) and the saturated specific humidity (q_s):

$$RH = \frac{q}{q_s} \cdot 100\% \tag{6}$$

The relative humidity (RH) is unchanged in the perturbed simulations, therefore the specific humidity changes in the same order as the saturated specific humidity, which is a function of the temperature given by the CC-relation shown in Equation 1 in section 2.3. Figure 9 shows the fractional change of the relative humidity and specific humidity, taken as the domain mean of the 10 LPS cases at the initial time in simulations. The relative humidity is unchanged, and the change in the specific humidity is 13 % (26 %) for the 2K (4K) simulations, i.e. 6.5%/K, in the lower part of the atmosphere.

To keep large-scale flow constant between the three simulations, the last modification includes an adjustment of the surface pressure (p_s), which mainly occurs over the high terrain. By starting with the hypsometric equation, an expression for the height of the terrain, Z_t, is obtained:

$$Z_t = \frac{R\bar{T}}{g_0} \cdot \ln\frac{p_{SL}}{p_s} \tag{7}$$

R is the gas constant for air, $\bar{T}$ is the average temperature in the layer, p_{SL} is the sea level pressur, and p_s is the surface pressure. When adding a temperature pertubation, the height of the terrain does not change, and by assuming that the sea level pressure does not change (e.g. Schär et al., 1996), the new surface pressure with the temperature pertubation can be

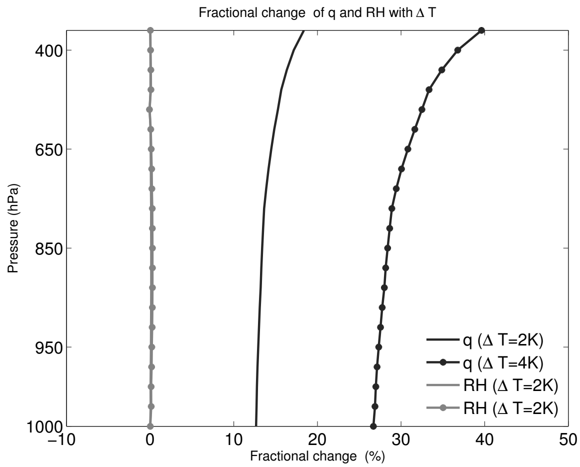

Figure 9: *The domain mean change in the relative humidity (RH) and specific humidity (q) between the control runs and the two perturbed runs (2K and 4K).*

calculated as:

$$p_{s,\Delta T} = \frac{p_{SL}}{\left[\frac{T}{(T+\Delta T)} \cdot \ln \frac{p_{SL}}{ps}\right]} \tag{8}$$

The new geopotential height (Φ) with the new temperature field and the adjusted surface pressure is calculated as:

$$\Phi_{\Delta T}(p) = \int_{p}^{p_{s,\Delta T}} R(T + \Delta T)\frac{dp}{p} \tag{9}$$

Where p is at some reference pressure level. The height of the surfaces of constant pressure has slightly increased by adding a temperature pertubation. The modification of the surface pressure ensures geopotential height distribution to be dynamically consistent with the new temperature field, and there are no changes in the large-scale gradients across the model domain.

4.3 Weather Research and Forecasting (WRF) model

The dynamical downscaling is performed with the regional model Weather Research and Forecasting (WRF) (Skamarock et al., 2008). WRF is a community model where the National Center for Atmospheric Research (NCAR) is leading the development. WRF is a fully compressible and non-hydrostatic atmospheric model, which is used for operational weather forecasting as well as climate research. The Advanced Research WRF (ARW) dynamical core, version 3.4.1 is used for the surrogate climate change experiments. The model and domain configurations have a 4km horizontal resolution and 50 vertical (eta) model levels, and the domain is run with 694 x 694 grid points. The domain covers India and parts of BoB, Himalaya and Arabian Sea. No convection parameterization was enabled for the 4km runs, since the convection was assumed to be explicitly simulated. The runs are performed with the Thompson microphysical scheme (Thompson et al., 2008), Yonsei University Planetary Boundary layer scheme (YSU) (Hong et al., 2006), Community Atmosphere Model's (CAM) longwave and shortwave radiation scheme (Collins et al., 2006), and the Noah Land Surface Model scheme (Niu et al., 2011). The model was run with SST update every 6 hour, and an alternative initialization of the inland lake's surface temperatures is included. This method uses the diurnal average temperature taken from the area close to the lake, instead of the SST from the closest ocean.

5 Summary of papers

This thesis consists of three papers, which are placed chronologically in the order they were written. All the papers are devoted to the monsoon low-pressure systems, where the associated precipitation is the main focus in the three studies, but each paper has a different perspective. In Paper I the dataset of the monsoon LPS is created, where statistical analysis on parameters that are important for the precipitation associated with the LPS in the ERA-Interim dataset is performed. In Paper II and III the results from the surrogate climate change simulations are studied: Paper II focus on changes in the parameters that are important for describing the LPS precipitation, while Paper III focus on the change in the extreme precipitation and runoff over the central Indian continent.

Paper I: The dynamic and thermodynamic structure of monsoon low-pressure systems during extreme rainfall events.

Silje Lund Sørland[1] and Asgeir Sorteberg[1,2]

1 Geophysical Institute, University of Bergen, Bergen, Norway.

2 Bjerknes Center for Climate Research, University of Bergen, Bergen, Norway

This was the first work done on the monsoon LPS in this thesis, and the aim was to generate a climatology of monsoon LPS that were related to an extreme rainfall event over the Indian continent. However, as discussed in the paper, to construct a climatology of a given weather phenomena is very sensitive to the underlying dataset and the criteria used to pick out the features. Thus a dataset of monsoon LPS, which had similar thermodynamical structures, i.e. cold core at lower levels and warm core aloft, was constructed. The LPS are detected in the ERA-Interim dataset, and the LPS that are generating an extreme rainfall event in both the ERA-Interim precipitation and the observationally based IMD rainfall dataset are selected for analyses. There were in total 39 LPS that fulfilled all the criteria during the time period 1979-2010. The main focus was on the investigations into which parameters that are important for the LPS precipitation, together with a discussion of the dynamic and thermodynamic structure of the LPS.

Key findings:

- The cyclone composite clearly shows the general structure of the LPS, with a pronounced cold core at lower levels and warm core aloft. Evaporative cooling from the falling precipitation is proposed to generate the cold core.
- The temperature gradients across the cyclone center are strongest in the early phase of the low. Based on this, it is suggested the baroclinic instability to be important in the development phase of the LPS, whereas the upward motion ahead of the low is maintained through latent heat release in the mature phase.
- From the composites of the time steps where the extreme precipitation is occurring, a colocation of the strong updraft and vertical velocity is shown. It is thought that the extreme rainfall events is a result of the LPS-dynamics, which is dominated by the CISK-mechanism at this stage of the low.
- Correlation and co-variability between the LPS precipitation and different meteorological parameters is performed, and the LPS precipita-

tion intensity shows a large sensitivity to the strength of the low-level velocity and specific humidity.

Paper II: The precipitation response in climate sensitivity experiments of monsoon low-pressure systems over India.

Silje Lund Sørland[1], Asgeir Sorteberg[1,2], Changhai Liu[3] and Roy Rasmussen[3]

1 Geophysical Institute, University of Bergen, Bergen, Norway.

2 Bjerknes Center for Climate Research, University of Bergen, Bergen, Norway

3 Research Applications Laboratory, National Center for Atmospheric Research, Boulder, USA.

In this paper 10 of the 39 LPS cases detected in Paper I is picked, and it is performed high-resolution climate sensitivity simulations, where all the temperature fields is perturbed with +2 and +4K. The specific humidity is correspondingly increased by 13 and 26 %. The mean changes between the control and perturbed runs following the LPS, is analyzed. The main focus is on parameters that are important for the modeled change in precipitation intensity.

Key findings:

- The precipitation associated with the LPS is increasing with up to 13%/K, which is twice the imposed initial moisture increase. This large increase in the precipitation is associated to feedbacks in the vertical velocities and the atmospheric stability.

- In the perturbed temperature simulations the LPS have a higher propagation speed and have developed into more intense storms. It is suggested that the storms in the warmer simulations have been intensified through the conditional instability of second kind (CISK) mechanism, since the condensational heating has increased together with the low-level convergence and vertical velocity, and as a result the depth of the surface low has strengthened.

Paper III: Low-pressure systems and extreme precipitation in central India: sensitivity to temperature changes.

Silje Lund Sørland[1] and Asgeir Sorteberg[1,2]

1 Geophysical Institute, University of Bergen, Bergen, Norway.

2 Bjerknes Center for Climate Research, University of Bergen, Bergen, Norway

This study is continuing the work from Paper II by using the same model outputs from the 10 simulations. However, here the focus is on the short duration extreme precipitation that is released over central India, instead of the mean precipitation following the low-pressure systems.

Key findings:

- The main difference between the control and perturbed runs is the change in the extreme precipitation. In the perturbed runs the intensity of the precipitation is increasing at all percentiles, and there are more frequent rainfall with very heavy intensities. This shift in the category of precipitation frequency and increase in the intensity, leads to a shift in which category that contribute most to the total precipitation: more of the precipitation is coming from the category with very heavy intensities.
- The runoff relative changes between control and perturbed runs in the different categories follow the precipitation changes, except the relative response of extreme intensity runoff, which for subdaily timescales are over twice the relative precipitation change.
- The mean precipitation change with warming in the central Indian region is up to 20%/K, and this precipitation change is larger than the precipitation responce found in Paper II. In Paper II the precipitation change following the low-pressure center was analyzed, but Paper III consider central India, where there are regions that have not been subject to precipitation in the control runs, but in the perturbed runs there have been precipitation, mainly due to an increased horizontal extent of the system and a slightly different propagation path.
- The LPS in the warmer runs are bringing more moisture further inland

that is being released as precipitation in the western region of central India, and as a result the precipitation change is largest further inland.

6 Discussion and future perspective

The 3 scientific papers presented in this thesis investigate the monsoon LPS and the associated precipitation. A well known tracking algorithm is used to detect the LPS in the ERA-Interim reanalysis dataset. LPS that are related to observed extreme rainfall events in India are picked out. 39 LPS cases fulfilled the criteria, and these cases were examined by using variables from the ERA-Interim reanalysis dataset. Next, 10 LPS cases were chosen and simulated with the high-resolution model WRF, using the surrogate climate change approach. The results were analyzed with respect to how the LPS precipitation could change in a warmer and more humid environment. The change in the mean precipitation following the LPS was investigated, as well as changes in extreme precipitation and runoff in the central Indian region. The main findings will be summarized in the following, sorted by the objectives given in the introduction.

How are the LPS that are connected to an observed extreme precipitation event represented in the ERA-Interim dataset?

The cyclone composites of variables from ERA-Interim reproduce the general characteristics of the LPS. The cold core at lower levels and warm core aloft is seen, and the composite also show a warm air sector ahead of the low and a cold air sector behind the low-pressure center. The updraft is located ahead of the surface low, collocated with the area of maximum precipitation. The LPS have a larger horizontal extent over ocean than over land, and during the time extreme precipitation occurs, the systems are more intense (stronger pressure gradients) and the precipitation cover a smaller horizontal area. Based on how the LPS are described in the literature, it seem as ERA-Interim reproduce the LPS dynamical and thermodynamical structure well (e.g. Godbole, 1977; Sikka, 2006; Krishnamurthy and Ajayamohan, 2010), however, the precipitation shows some deficiencies. The ERA-Interim precipitation was compared with the TRMM precipitation, and ERA-Interim produce too much precipitation each time step, and the precipitation is not placed on the correct place. Moreover, when picking out the extreme rainfall events that were common in both ERA-Interim and the observational based IMD dataset, it was found that ERA-Interim underestimates the extremes and the extremes often develop in the wrong place/at the wrong time.

Which meteorological parameters are important for understanding the variability in the precipitation intensity associated with the monsoon LPS?

Based on a simple description of the precipitation intensity, the precipitation is related to the atmospheric temperature, atmospheric stability, vertical velocity and the specific humidity. The co-variability between the LPS precipitation and different variables was investigated, and it was found that a large part of the precipitation variability could be explained with only a small set of parameters. The upward motion and specific humidity at 750hPa shows the largest correlation with the precipitation. The temperature at 750hPa shows no significant correlation with the LPS precipitation, and the lack of correlation may be due to a balance between the cooling from adiabatic ascent and heating from condensation. Because of evaporative cooling from falling precipitation, the temperature at 950hPa is negatively correlated with precipitation. The atmospheric stability, given as the difference between the temperature at 750 and 950hPa, show a weak correlation with the precipitation. It seems as the lower atmospheric temperature, which is influenced by evaporative cooling, mainly controls the variability in the temperature difference.

How sensitive is the average precipitation associated with the LPS to an increase in the atmospheric temperature and moisture content?

The mean precipitation following the monsoon LPS increases with up to 13.2 %/K in the climate sensitivity simulations, which is twice the increase in the atmospheric moisture content expected from the CC-relation. This large increase is explained by the imposed specific humidity increase, a dynamic feedback giving stronger upward motions and a thermodynamic feedback decreasing the atmospheric stability. The precipitation response can be interpreted by a feedback loop: An increase in the temperature increases the atmospheric moisture content. More moisture leads to an increase in the condensation and latent heat. The vertical velocity and atmospheric stability is influenced by the increase in the latent heat, and is therefore amplifying the precipitation response.

How sensitive is the extreme precipitation generated by LPS and associated runoff in the central India region to a warmer and more humid atmosphere?

Comparing the warmer model runs to the control runs, it is found that the LPS are covering a larger area, propagating faster and able to bring more moisture across the Indian continent, which is released as precipitation further inland. The warmer runs show that in the central Indian region the precipitation intensity is increasing at all percentiles, and an increase in the frequency of moderate, heavy and very heavy precipitation events occurs at the expense of low precipitation events. More of the precipitation is coming from the category with very heavy intensities, while there is a decrease in the contribution from the other categories. The runoff shows similar response as the precipitation to temperature warming, except for the changes in the extreme intensities. The relative change in the 99th percentile intensity for the runoff is twice the relative change in the precipitation intensity, but for lower percentiles the relative change is similar for the precipitation and runoff. Based on this, it is suggested that there is an increased risk of more severe flooding events in the central India region during the passage of these LPS in a warmer climate. It can also be speculated that there may be severe flooding in regions that has been less prone to flooding before, since the storm is able to bring high-moisture content further inland.

Main findings and perspective

The main motivation for this thesis was to understand how sensitive the LPS precipitation is to a warming world. The results clearly show that in a warmer and more humid atmosphere, the LPS can produce more precipitation, precipitate with a higher precipitation rate and also release precipitation further into the Indian continent. If this is how the LPS may develop in a warmer atmosphere, the impact from the LPS precipitation on the Indian society can be large. However, it should be emphasized that these experiments are idealized, and assumes the large-scale flow to be unchanged. In a warming climate the LPS would develop on a different monsoon background flow. Thus it cannot be expected that the lows will behave in the same manner in a warmer climate, since the large-scale dynamic will influence the propagation and frequency of the LPS.

The LPS is such an important feature of the monsoon circulation, and account for up to 50 % of the monsoon precipitation (Yoon and Chen, 2005), therefore it is important to investigate whether or not the frequency or intensity of the monsoon LPS will change with atmospheric warming. There are several challenges associated with the LPS research, related to the uncertainty regarding the climatology of the monsoon LPS, whether there has been a negative trend or not of the monsoon depressions. Cohen and Boos (2014) propose that there are deficiencies in the monsoon LPS dataset constructed by the Indian Meteorological Department (IMD), since the declining trend of monsoon depressions seen in IMD depression time series, is not found in other time series constructed with different underlying datasets and with different automated detection methods. Difficulties in constructing climatology of monsoon LPS with an automated tracking method was also found in Paper I, where the number of detected features was very sensitive to the different criteria used to filter out the tracking errors. To determine the choice of objective criteria when using automated detection methods, it is important to know the structure of the feature to be detected. Since we decided to use the thermal structure as one of the criteria, we have excluded LPS that do not have a clear cold core at lower levels and warm core aloft. However, this criterion may have removed dry vortices, since the cold core is shown to be a result of evaporative cooling, in addition to have excluded vortices that may have tropical cyclone intensity, which have warm core throughout their vertical extent.

To assess how the precipitation may change with climate is an important but challenging task. The tropical precipitation, which is mostly of convective character, should be simulated with a very high horizontal resolution, where convection is explicitly resolved. Moreover, to know how good your model is, you need to evaluate your modeled precipitation against observations. Due to the heterogeneity of the precipitation, and the limitations of observations in the tropics, it is difficult to obtain long time-series of high-resolution precipitation datasets. In India there is a long history of measuring the precipitation due to the importance the monsoon progress has on the society. The IMD rainfall dataset covers a long time period, and the information from the rain gauge are interpolated into gridded data. The number of stations within each grid square and the interpolation method will affect the quality of the precipitation, and especially the representation of the extremes will be affected. We evaluated the WRF precipitation against the IMD precipitation in the central Indian region, and found the WRF precipitation to simulate the IMD precipitation fairly well. One advantage of performing a sensitivity experiments as we have in this thesis, is that the aim is not to simulate the precipitation perfectly, but we are focusing on the differences in the precipitation between the control and perturbed simulations.

Several studies have investigated how the precipitation scales with temperature warming (Allen and Ingram, 2002; O'Gorman and Schneider, 2009; Romps, 2011; Loriaux et al., 2013). In this thesis we performed multiple linear regression to understand the model results, and used the regression coefficients to assess the sensitivity of changes in the precipitation to changes in any of the predictors. The results have to be interpreted with caution, since the predictors are not independent. However, we got a good estimate of which parameters that are important for the precipitation variability. In Paper II we clearly show that the change in precipitation with temperature warming do not scale as the CC-relation. The changes in the atmospheric stability and vertical velocity are also important, where the latter has the largest impact on the LPS precipitation. With this we want to highlight the different feedbacks affecting the precipitation intensity with an increase in the temperature. However, to properly assess how the precipitation associated with the LPS may change, climate models has to be able to capture the monsoon LPS (e.g. Turner and Annamalai, 2012), so that LPS can be properly simulated in the future background flow. Only then can we fully assess how the LPS precipitation may change in a future scenario.

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