TOWARDS AN IMPROVED UNDERSTANDING OF REGIONAL SCALE CLIMATE CHANGE IN THE NEPAL HIMALAYAS

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RUDRA KUMAR SHRESTHA

SCHOOL OF EARTH, ATMOSPHERIC AND ENVIRONMENTAL SCIENCE
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ABSTRACT

The effects of enhanced greenhouse gas concentrations on Earth’s climate are well understood. However, the impacts of anthropogenic aerosol particles, in particular due to the many aerosol-cloud indirect feedback mechanisms are not fully or even explicitly quantified as yet. This PhD seeks to contribute to improve our knowledge and understanding of aerosol – precipitation interactions over the Nepal Himalayas region and their consequences for precipitation patterns there. The research was carried out using the cloud-resolving Weather Research and Forecasting (WRF) model through a series of sensitivity studies and supported by literature reviews of satellite and field observations, although the latter are sparse.

To complement the modelling studies, from March to December 2011, aerosols and surface meteorology were also continuously measured at Nagarkot (Lat: 27.7°N, Lon: 85.5° E, Alt: 1900m), Nepal, located in the eastern flank of a bowl shaped Kathmandu valley. The location was chosen to provide a representative vertical profile of aerosol and the impact on topographical flows. Our results showed a unique pattern of diurnal pollution circulation within the valley with a morning and evening peak. The evening peak, which is higher than the morning peak is attributed to the light wind blowing through the valley carrying locally generated fresh evening pollution, further enhanced by recirculations of aged pollutants through suppression of the mixing layers as suggested by a previous study at a different location. The morning peak is caused by calm wind conditions followed by the transitional growth of the nocturnal boundary layer. It is found that the thermally driven mountain – valley wind circulations are responsible for ventilation of pollutants.

The WRF simulations showed that a sophisticated double moment bulk microphysics parameterization scheme performed best, which did not show any statistically significant difference compared to the observed data at 80% confidence interval using a Chi-squared goodness of best fit test. A sensitivity analysis of aerosol and temperature perturbations on the monsoon precipitation was conducted. We found that the model represented the first indirect effect reasonably well however, rainfall was not particularly sensitive to the aerosol perturbations used, due to the poorly documented role of the ice phase processes which assume a greater importance in this region due to the influence of topography and diurnal heating cycle. Further model studies focusing on chemical properties of aerosol and sensitivity of Ice Nuclei (IN) to precipitation in this region are recommended. In contrast, the effects of temperature perturbation were found to be significant, more so than the currently modelled aerosol indirect effects, suggesting that reduced frequency but intense rain events are likely over the Himalayas as the climate warms.
DECLARATION

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Faculty: Engineering and Physical Sciences
Thesis Title: Towards an improved understanding of regional scale climate change in the Nepal Himalayas

Declaration to be completed by the candidate:
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CHAPTER 1: Introduction

Indian Monsoon precipitation is susceptible to climate change and air pollution, which serves billions of people living across south Asia. Understanding the mechanisms by which the system is affected plays a key role to reduce the uncertainty in our projections of future monsoon precipitation over the topographically complex region.

1.1 Motivation

The Himalayas is known as the ‘third pole’ (Lami et al., 2010) as it has a huge amount of ice reserves and most of the topography is not easily accessible. It is also termed as the ‘water tower’ because the Himalayan glaciers feed Asia’s major rivers such as the Ganges, Indus, Brahmaputra, Salween, Mekong, Yangtze and Huang Ho, serving more than 1 billion people living in the downstream (Singh et al., 2011). However, very little is known about the region, to give one example, the Intergovernmental Panel on Climate Change (IPCC), a leading international body for the assessment of climate change, published a wrong and unreliable information about melting of the Himalayan glaciers, which was cited from a non-peer reviewed literature on its Fourth Assessment Report (AR4) in 2007 (Schiermeier, 2010).

Monsoon floods and droughts are major threats to the people living in the South Asia. The region receives up to 80% of annual rainfall during the summer season (Shrestha, 2000; Turner and Annamalai, 2012), which often cause floods and landslides. However, a long monsoon break can contribute to a sever drought condition (Krishnan et al., 2000) that is not uncommon in the region. A small change in rainfall patterns and/or a few days delay in the arrival of the monsoon can have tremendous impact on agriculture, economy, water resources and ecosystem, which could pose a serious threat to sustainability of the region. Hence, accurate forecasting of the Indian monsoon is a major priority due its
wide-ranging potential impacts on agrarian based society ranges from individual farmer level to large scale agribusiness for export.

Importance of the Indian monsoon is not only limited to agriculture however, there are several other sectors which depend on the monsoon rain. South Asia considerably depends on hydro-based electricity generation. Majority of the existing infrastructures are designed for the run-of-the-river type, where no water storage facility is provided that may have serious implication in the future due to high energy demand attributed to rapid industrialization, population growth and climate change. The region is already facing a severe power deficit, for example, in Nepal power outage occurs throughout the year, its duration could reach up to 18 hours a day in the dry season and the situation will become even worse in the future. Nepal has been a popular tourist destination over the decades, the spectacular Himalayas attract tourists from all over the world, however due to poor and/or no forecasting of weather many mountaineers and trekkers have lost their life during the journey. Furthermore, aviation safety is also an important issue, number of accidents have been reported due to bad weather condition. Therefore accurate forecasting of weather is necessary from household to business and policy level in order to plan their projects in well advance.

The study region is located in between two rapidly industrialized countries in the world, India and China, which are estimated to contribute a major part of the global greenhouse gases emission and air pollutants. Unlike the Greenhouse gases, aerosols exert warming and/or cooling effects on the atmosphere. Furthermore, the aerosol particles modulate cloud microphysics because it can serve as clouds nuclei, also deteriorate the ambient air quality and these effects are heterogeneous in nature. Such processes are even more complex and hard to predict over the Himalayas where advected aerosols from the Indo-Gangetic plains are thought to be trapped in the valleys subsequently modulates their properties and thereby formation and distribution of orographic clouds and precipitation (Shrestha et al., 2010).

Satellite, aircraft, and ground-based observations and model simulations show that air pollution can be transported over long distances throughout the
global atmosphere, e.g., from the Sahara desert to the Himalaya (Carrico et al., 2003), eastern Asia to the western U.S. (Berntsen et al., 1999) and from North America to Europe (Stohl and Trickl, 1999). Visibility degradation and the Atmospheric Brown Cloud (ABC) phenomenon are already prominent environmental issues across South Asia (Ramanathan et al., 2005). Furthermore, the role of Black Carbon (BC) in the melting of Himalayan glaciers is also spotlighted (Ramanathan and Carmichael, 2008).

Global warming attributed to the increased greenhouse gases is projected to increase the Earth’s mean temperature (Forster et al., 2007), the effects sustain for a long period of time in the atmosphere. The enhanced temperature increases moisture holding capacity of the air (Soden et al., 2005), which could lead to potential for extreme weather events. Furthermore, temperatures over the Mountainous region are increasing faster than in the other part of the world (Liu and Chen, 2000; Shrestha et al., 1999). The effects are already observed in the Himalayas because accelerated retreating of glaciers and number of glacier lake outburst have been reported in the recent years (Jain et al., 2012).

Future population growth will inevitably have large-scale impacts on urban air pollution. According to latest UN report 6 billion world’s current population will be expected to reach 9 billion by 2050, led by a doubling of the number of urban dwellers in less developed regions (UN, 2012). Many cities in developing countries such as Nepal already experience serious air quality problems and these cities are likely to benefit more slowly from emission-reducing technologies.

The Department of Hydrology and Meteorology (DHM), Nepal, a governmental agency, is solely responsible for weather forecasting across the country. The DHM is not fully equipped with currently available sophisticated computing facility, which has very few local exports and still uses ‘stone edge’ technology to predict the weather (Rosoff, 2003). However, to the best of our knowledge, very recently they have started to use a coarse resolution Weather Research and Forecasting (WRF) model to forecast rainfall across south Asia.
Again these coarse resolution (9 km) simulations can not provide very detail features of the weather.

Although there are few past research on air pollution modelling that focus on the Kathmandu valley’s circulation (Kitada and Regmi, 2003; Panday et al., 2009; Regmi et al., 2003), its interaction with meteorology has not been well documented yet. Scientific understanding of such processes is limited by resource constraint as the money often competes with other basic needs. A field experiment described in this thesis, which is a collaborative work with a local research institute in Nepal, provides an important database that can be used not only to verify modelling results but also guides to formulate pollution abatement policy and mitigation measures. As the country does not have effective emission control policy and enforcement mechanisms, the PhD thesis would be highly valuable towards contribution to science and to the public in Nepal.

The overall aim of this thesis is directed to improve our understanding of the Indian monsoon system and its interactions with aerosol pollutants under the plausible climate change conditions in the Nepal Himalayas which is exacerbated by extremes in topography. The research involves ground-based measurement and supported by model simulations. We address how the precipitation formation processes are affected by the choice and assumption of microphysical treatment in the numerical model. Another aims of this research is to develop climate change database through networking and collaboration with various scientific institutes in Nepal (e.g. DHM, ICIMOD and IOE) and also to make a closer link between the United Kingdom and Nepalese scientific communities.

1.2 Thesis organization

This thesis is prepared in the alternative format as permitted by the University of Manchester thesis submission guidelines. The thesis contains three journal papers. The first paper, Seasonal and diurnal variations of meteorology and aerosol concentrations at a mid-hill station in the foothills of the Nepal Himalayas, Nagarkot (1,900 m asl), uses observed aerosol number concentration
and size distribution, and meteorological data, which was collected during our field observation from 2011 to 2012. This paper provides general idea of the aerosol characteristics across the region, which is an important input for the modelling study. The second paper, sensitivity of WRF cloud microphysics to simulate a convective storm in the region of complex terrain, focus on the simulations of a convective event that evolved across central Nepal and also compare the results from different microphysical parameterization schemes. The objective of this paper is to analyze performance evaluation of different microphysical schemes available within the WRF model. The third paper, sensitivity of Indian monsoon precipitation to aerosols and temperature perturbations over the foothills of the Nepal Himalayas, investigates sensitivity of temperature and aerosol perturbations to formation and distribution of clouds and precipitation. The third paper has been submitted (under review) to Atmospheric Research for publication and the first and second papers are being submitted in Science of the Total Environment and Annales Geophysicae respectively.

The thesis is organized into six parts: general introduction, literature review, seasonal and diurnal variation of aerosols and meteorology, sensitivity of convective storm to WRF microphysics, sensitivity of precipitation to aerosols and temperature perturbation and general conclusions.

1.3 Reference


CHAPTER 2: Literature Review

This chapter will review the current scientific understanding of the Indian monsoon system. We will mainly focus on driving factors of the monsoon processes and their inter-linkage to other regional and global phenomenon. We will also attempt to explain the observed variability of meteorological and aerosol components over the Nepal Himalayas using satellite measurements and related albeit limited previous surface observational and modelling studies.

2.1 Overview of Indian Monsoon

The monsoon is a large scale seasonal reversal of prevailing regional wind speed and direction. These changes in wind patterns are associated with the seasonal variation of solar insolation. A strong contrast between temperature over land and ocean surfaces is formed due to differential surface heating which leads to the generation of strong winds from the ocean to the land. This widespread phenomenon occurs across south Asia is referred to generally as the Indian monsoon, but similar phenomena are also observed in the other parts of the globe, in West Africa and Australia. These lead to very large changes in regional precipitation regimes, characterising a shift from very dry to very wet seasons.

The Indian monsoon mechanism was first proposed by Edmund Halley in 1628. Webster et al. (1998) also championed the explanation that the thermal contrast between land and ocean was a key driving factor causing the monsoon precipitation. Simpson (1921) supported Halley’s hypothesis but also highlighted the role of the giant Himalayan barrier, which forces the monsoon current to ascend and in turn condense and release latent heat that provides a pseudo self-sustaining energy source for the monsoon system overall (Cadet, 1979).

During the boreal summer, a significant warming is observed over the Himalaya – Tibetan Plateau region due to increased sensible and latent heat fluxes, leading to a strong temperature gradient along meridional and zonal
direction, as shown in Figure 2.1, derived from 30 years of NCEP/NCAR reanalysis data. The boreal winter (December – February) shows different patterns of temperature with low values over the Tibetan plateau and high values over the south Indian Ocean and Australia. Changes in the large scale wind patterns, which flow in the forward and reverse directions, as shown in Figure 2.2, are caused by this seasonal variation in temperature gradient that also brings with them a change in precipitation regime as shown in Figure 2.3.

**Figure 2.1:** Seasonal temperature variation in the mid – troposphere (500 hPa) in the NCEP/NCAR reanalysis dataset (1981 – 2010). Unites are in °C.

It is generally thought that there is a coupling between the Indian monsoon and El Nino and southern oscillation (ENSO), which may cause inter-annual variability of the monsoon precipitation (Turner and Annamalai, 2012). The ENSO mechanism, attributed to deceleration of the trade winds, caused by the low equatorial pressure gradient between the Western and Eastern Pacific regions, enhances convection over the central Pacific and affects movement of the
monsoon winds within the Indian subcontinent. These in turn suppress convection and precipitation across the region (Turner, 2005). An analysis of 132 years of observations (Kumar et al., 2006), showed that severe droughts over India normally coincide with El Nino events. Studies of the Nepal region, Shrestha (2000), suggest that inter-annual variability of precipitation across the country is strongly correlated to the Southern Oscillation. However, Gadgil (2003), have questioned the degree to which ENSO actually influences the Indian monsoon precipitation.

Figure 2.2: Seasonal reversal of low-level (925 hPa) winds shown for the boreal summer and winter seasons in the NCEP/NCAR reanalysis dataset (1981 – 2010). Indian monsoon region is highlighted using a rectangle.
Modelling of the monsoon under future climate change scenarios is a major challenge for climate scientists because the current climate models do not represent the distribution of monsoon rainfall well at a regional scale (Turner and Annamalai, 2012). In a review of the Indian monsoon Turner and Annamalai (2012) pointed out that multi-model mean simulations of the summer monsoon precipitation (in the doubling of CO$_2$ concentration scenarios), predicted enhanced rainfall over the Himalayas, in contrast to observations which show a decreasing trend (Bollasina et al., 2011; Duan et al., 2006). These uncertainties in predicted rainfall may have been associated with the enhanced anthropogenic aerosols across the region, which reduces circulations over Indian Ocean and subcontinent (Bollasina et al., 2011). These aerosol feedbacks mechanisms are not well simulated at present.

Figure 2.3 Seasonal variation of surface precipitation rate (mm/day) in the NCEP/NCAR reanalysis (1981 – 2010).
Climatologically the onset of the Indian monsoon across central Nepal generally occurs in the second week of June (Barros and Lang, 2003; Bollasina et al., 2002; Bonasoni et al., 2008), as depicted in the Figure 2.4, and projected by the Indian Meteorological Department (IMD). Li and Yanai (1996) and Xavier et al. (2007) found that the onset date of the monsoon is consistent with reversal of the meridional temperature gradient in the upper troposphere in the region south of the Tibetan Plateau. This temperature gradient is associated with the surface warming over the Tibetan Plateau. However, this process is weaker over the ocean due to latent heating and subsequent adiabatic cooling of the air parcel on ascent.

On a synoptic scale the monsoon onset period is characterized by an increase in water vapour content and convective instability over the Himalayas.
(Barros and Lang, 2003; Yasunari, 1976). This is accompanied by a weakening of upper level westerly flow, which is consistent with the onset of the monsoon depression over the Bay of Bengal and its subsequent movement towards the Himalayas and which transports monsoon moisture over the region (Barros and Lang, 2003). Uneo et al. (2008) and Kurosaki and Kimura (2002) also investigated onset of monsoon in the Himalayas.

2.2 Distribution of rainfall in Nepal

Rainfall over the Himalaya region is dominated by the southwest monsoon, which generally occurs during the summer season (June – October), and contributes about 80% of the total annual rainfall there (ICIMOD, 1996; Shrestha et al., 2000) but with significant annual variability. Higuchi et al. (1982) carried out precipitation observations over 3 summer monsoon seasons (1974, 1976 and 1978) along the South – North section of a river basin in eastern Nepal. The observed total precipitation amount over the peaks in the region was found to be several orders of magnitude greater than in the valley basin. Higuchi et al. (1982) and Ageta (1976) indicated that the frequency and amount of precipitation over these peaks are much more pronounced during the day time, which they attributed to cumulus convection. In contrast, the valley basins received significantly more precipitation during the evening to midnight period. They proposed that this interesting nocturnal maximum precipitation phenomenon was caused by a convergence of weak up-valley winds, which remained connected to the basin flow, and which may have been caused by the latent heat release following cloud formation and precipitation (Ohata et al., 1981), and the katabatic winds associated with radiative cooling in the foothills of the Himalayas.

Barros and Lang (2003) also reported this nocturnal maximum precipitation phenomenon during their monsoon monitoring project over central Nepal. In contrast to the previously proposed hypothesis, they argued that this phenomenon is significant mainly during the summer monsoon season. A steady south-westerly moist monsoonal flow occurs throughout the season, which is further enhanced by up-slope and up-valley flows during the day time causing
precipitation over the ridges and peaks. However, during the night time down-slope and down-valley winds interacts with the monsoon flow causing convergences over the valley basin producing maximum nocturnal precipitation across the region. Hence, these observations suggest that diurnal monsoon precipitations over the Himalaya region are driven more by coupling of local circulations with the larger-scale monsoon flow.

Figure 2.5: Spatial distribution of mean annual precipitation, adopted from (ICIMOD, 1996).

An analysis of the spatial distribution of mean monthly precipitation across Nepal was carried out using rainfall records from 264 stations available for a 5 year period (ICIMOD, 1996). The mean annual precipitation was computed using the 12 mean monthly precipitation values from these stations (Figure 2.5). This reveals that the region defined by Pokhara city and its immediate surrounds, which lie on the wind ward side of the mountain range in central Nepal, maximum annual precipitation occurs with values up to 5000 mm being recorded. In contrast, the Jomsom area, located north of Pokhara city, and on the lee ward side of the mountains, receives the minimum annual precipitation (typically 100 mm). During winter season the country receives its minimum precipitation, as shown in
Figure 2.6, whereas the heaviest precipitation occurs during the summer monsoon season, as suggested by Figure 2.7.

Figure 2.6: Same as Figure 2.5 but for January (winter season).

Figure 2.7: Same as Figure 2.5 but for July (monsoon season).
2.3 Distribution of temperature in Nepal

Near surface air temperature is generally considered a useful, but sometimes problematic, indicator of climate change (Kumar and Hingane, 1988), with a small change in regional air temperature might lead to a tremendous impact on atmospheric circulation, in turn leading to significant changes in hydrological cycle. A number of model studies have been conducted to investigate the influence of temperature trend changes on global, regional and local scales. Shrestha et al. (1999) examined temperature trends over Nepal using 23 years of observational data (1971 – 1994) from 49 measuring stations across the country. They found that warming trends over the high altitude region (0.06°C to 0.12°C per year) were considerably higher than the mid-hill and ‘Terai’ (southern part of the country, altitude <200m) region (0.03°C per year). Shrestha et al. (1999) also indicated that average all-Nepal trends will likely follow similar warming trends predicted for the Northern Hemisphere suggesting a coupling between global and regional climate phenomenon, however the magnitude of the trend would be enhanced over the Himalayas.

High resolution meteorological and other climate relevant atmospheric observations are sparse in Nepal (Fujita et al., 2006; Grabs and Pokharel, 1992), mainly due to financial and topographical limitations and as well as a lack of local experts to operate monitoring stations, as discussed by (ICIMOD, 1996). This is preventing improved understanding of climate change in one of the potentially most sensitive regions of the world. Furthermore, meteorological data from the important Tibetan Plateau region are also very limited (Liu and Chen, 2000; Shrestha et al., 1999). However, there are now a growing number of studies across India which provide us with useful insights to approximate expected meteorological trends over the adjoining Nepal region. Hingane et al. (1985) carried out a long term trend analysis of surface temperature in India using 81 years data (1901 – 1982) measured across the entire continent. They found a linear warming trend for mean annual temperature of 0.4°C/100 years, and a more pronounced trend, 0.5°C/100 years to 0.6°C/100 year, was observed in the northern part of the country. An accelerated warming trend (0.22°C per decade)
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was observed over all of India during last quarter of the 20th century (Kothawale and Kumar, 2005).

An analysis of temperature trends over the Tibetan Plateau during the 41 year period from 1955 to 1996 using records from 97 stations reveal a mean annual warming trend of 0.16°C per decade. A more pronounced warming trend (0.32°C per decade) was observed in the winter season and there was a tendency for these warming trends to increase with elevation (Liu and Chen, 2000). Furthermore, Lie and Chen (2000), also mention that the region is most sensitive to predicted global climate change, because the warming trends over the Plateau is higher than in the Northern Hemisphere compared to other similar latitudinal zones in other parts of the world.

The International Centre for Integrated Mountain and Development (ICIMOD) carried out a spatial distribution analysis of surface temperature available from 80 stations, covering a period of approximately 5 years, across Nepal (ICIMOD, 1996). Regression analysis was used to calculate maximum and minimum monthly temperatures, where the effect of elevation, latitude and longitude were considered. Figure 2.8 shows the mean maximum temperature for June, one of the hottest months of the year, which indicates that more than 2.5% of the country experiences warmer temperature ranges from 36°C to 39°C. Similarly, mean minimum temperatures, in January (Figure 2.9), shows that more than 28% of the country, mostly in the high Himalayan region, experiences very cold weather, sometimes below -3°C.
**Figure 2.8:** Spatial distribution of mean maximum temperature for June, adopted from (ICIMOD, 1996).

**Figure 2.9:** Same as Figure 2.8 but for mean minimum temperature for January.
2.4 Variability of Aerosol Optical Depth in Nepal

The Aerosol Optical Depth (AOD) parameter is defined as the ‘amount of attenuation of direct solar radiation passing through the atmosphere by scattering and absorption due to aerosol’ (Kumar et al., 2009), so it is often used as a quantitative measurement of extinction of solar radiation between the point of observation and the top of the atmosphere. AOD will depend on the amount, size distribution and chemical composition of aerosol within the atmospheric column (Gadhavi, 2005). AOD is one of the key aerosols optical properties used to describe the influence of aerosol loading and its interaction with radiation and is used for evaluating radiative forcing in climate modelling studies.

Spectral analysis of AOD provides important information to identify aerosol characteristics such as particle size distribution (Kumar et al., 2009). However, data scarcity in the Himalaya region limits our understanding of net aerosol properties and their effects on its regional climate, although is changing. Until recently satellite derived data have been relied on to improve characterization of aerosol properties and their climate influence across the region.

A significant temporal and spatial variation of AOD is observed over India as suggested by MODIS satellite images (Gadhavi, 2005) and a similar seasonal variation of AOD is found over the Nepal region (Figure 2.10). A higher value of AOD in the pre-monsoon season compared to a lower AOD in the post-monsoon season is observed over the Nepal region. Kumar et al. (2009) also found a significant seasonal variation of AOD over Southern India, with higher AOD during pre-monsoon and lower in the monsoon season. Bonasoni et al. (2008) analyzed AOD in the higher altitude region of Nepal at Khumbu (5079 m asl), in the Khumbu region of Nepal, over the period March to October 2006. They found higher AOD (in the 500nm wavelength region) in August with March showing the smallest AOD values.
Figure 2.10: Satellite based observation of seasonal variation of aerosol optical depth over the Nepal Himalaya and foothills a) winter, b) pre-monsoon, c) monsoon, and d) post-monsoon seasons. Analyses and visualizations used here were produced with the Giovanni online data system, developed and maintained by the NASA GES DISC.
2.5 Aerosol direct and indirect effects

Aerosol particles affect the climate system via a variety of direct and indirect feedback mechanisms (Albrecht, 1989). First, they scatter and absorb solar radiation. Second, they scatter, absorb and emit thermal radiation, which in turn alters the radiation fluxes at the surface and the top of the atmosphere, referred to as direct effects (Forster et al., 2007; Lohmann and Feichter, 2005; Ramanathan and Ramana, 2003; Schwartz, 2008). Radiative forcing associated with the direct effects of aerosols is known as direct forcing.

Figure 2.11: Schematic diagram showing aerosols forcing adopted from Schwartz (2008).

Figure 2.11 shows a schematic diagram of several feedback mechanisms by which aerosol can influence the formation of cloud and precipitation as well as the lifetime of those clouds. The small black dots represent aerosol particles, the larger open circles, clouds droplets, and the straight lines represent the incident and reflected solar radiation. The wavy lines represent longwave radiation and the grey lines represent rainfall. The length of the lines represents the possible magnitude of the different fluxes. The unperturbed cloud contains larger aerosol particles as these forms on naturally occurring aerosols. On the other hand, the perturbed cloud contains high concentrations of small particles, emitted from both natural and anthropogenic sources. Sulphate, organic carbon, black carbon and mineral dust are identified as major anthropogenic components that exert significant direct radiative forcing on climate (Forster et al., 2007). Among the
anthropogenic components dust and black carbon absorb considerable amounts of solar radiation, which can have a strong global warming potential. Over the Himalayas black carbon aerosols are thought to have a similar global warming potential as CO$_2$ (Ramanathan and Carmichael, 2008).

The indirect effects of aerosol arise due to mechanisms by which aerosols modify the optical properties, the amount and lifetime of clouds (Forster et al., 2007), also known generally as aerosol – cloud interactions. As depicted in Figure 2.11, the so-called first indirect effect, also known as the cloud – albedo or Twomey effect, states that an increase in cloud albedo, would be caused by an increase in CCN concentration for a fixed amount of liquid water content (Twomey, 1977). This hypothesis has now been supported by number of observational studies. Rosenfeld (1999) analyzed the impact of enhanced aerosol emissions from forest fires on tropical rainfall over Indonesia using the Tropical Rainfall Measuring Mission (TRMM) satellite and found a reduced cloud droplet effective radius in the smoke rich zone. Satellite studies of marine stratocumulus show that increases in CCN concentration, due e.g. to aerosol plumes from ship emissions, decreases droplet size resulting in the clouds becoming more optically reflective (Coakley et al., 1987). Due to increases in CCN concentrations, enhanced cloud droplet number concentrations and smaller droplet effective radii were also observed over the Indo-Gangetic Plain (IGP) during the CAIPEEX experiment (Cloud Aerosol Interaction and Precipitation Enhancement Experiment) (Konwar et al., 2012).

The second indirect effect, also known as the cloud – lifetime or “Albrecht effect” arises because a reduced average cloud droplet size suppresses precipitation formation efficiency, tending to increase liquid water content, cloud lifetime (Albrecht, 1989), and the cloud thickness (Pincus and Baker, 1994). Albrecht (1989) proposed that the smaller cloud droplets induced by the Twomey effect suppress droplet coalescence leading to reduced precipitation efficiency. Lohmann and Feichter (2005) also support the Albrecht hypothesis that the smaller droplets formed in the mixed-phase clouds suppress the riming process and reduce precipitation. However, observation from a particular region or cloud type is not always consistent with theoretical understanding or cloud that form in
a different region or a different type of cloud. The Albrecht hypothesis was based on observations of sub-tropical marine boundary layer clouds, and in contrast, Connolly et al. (2006), suggest a different view in case of the deeper tropical convective clouds, where enhanced aerosol concentration may accelerate the riming process, which in turn may produce more precipitation during the dissipation state of a convective storm, showing little change in the total precipitation amount. Such contradicting results suggest that the aerosol indirect effects are very uncertain due to non-linear feedbacks (Lohmann and Feichter, 1997) and will be case and region dependent. The aerosol indirect effects are also discussed in chapter 5 (i.e. monsoon modelling paper) of this thesis as part of the detailed model simulation study.

2.6 Reference


CHAPTER 3: Seasonal and diurnal variations of meteorology and aerosol concentrations at a mid-hill station in the foothills of the Nepal Himalayas, Nagarkot (1,900 m asl)

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Authors: Rudra K. Shrestha\textsuperscript{1,2}, Martin W. Gallagher\textsuperscript{1}, Paul J. Connolly\textsuperscript{1}

\textsuperscript{1}Center for Atmospheric Science, School of Earth, Atmospheric and Environmental Sciences (SEAES), The University of Manchester
\textsuperscript{2}Sustainable Consumption Institute (SCI), The University of Manchester

Contribution from others: Martin Gallagher helped with handling the equipments (GRIMM optical particle counter and Vaisala weather sensor) before going for the field work and data analysis and also reviewed and commented on the paper. Paul Connolly wrote some of the scripts to read the GRIMM data. All work was carried out under the supervision of Martin Gallagher and Paul Connolly.

Abstract

A 10-months (March – December 2011) long monitoring experiment to investigate seasonal and diurnal variation of aerosol size distribution at a site on the eastern edge of the Kathmandu valley (Nagarkot) at an altitude of 1,900m asl was carried out as part of a study on katabatic and anabatic influence on pollution dispersion. Aerosols and meteorological variables were recorded by an optical particle counter (GRIMM Model 1.109, size range 0.25-32 \(\mu\)m) and a Vaisala weather sensor respectively. The seasonal mean showed total aerosol number concentration was highest during the post-monsoon season (919±694cm\(^{-3}\)) followed by pre-monsoon (746±561cm\(^{-3}\)) and monsoon (374±361cm\(^{-3}\)) periods. In
general it was found that the fine particle number concentration 
\((0.25\mu m \leq D_p \leq 2.5\mu m)\) dominated in all seasons, however, contribution by coarse 
particles \((D_p > 2.5\mu m)\) was more significant in the monsoon season.

The mid-hill results show a regular diurnal pattern of pollutant circulation 
in the valley with a morning and an evening peak. The daily twin peaks are 
attributed to calm conditions followed by transitional growth and break down of 
the valley boundary layer. The peaks are generally associated with enhancement 
of the coarse particle fraction. The evening peak is generally higher than the 
morning peak, and is caused by the fresh evening pollution from the valley from 
increased local activities and re-circulation of trapped pollutants through 
suppression of the mixing layer. The morning peak is caused by growth of the 
mixing layer. Moreover, relatively clean air masses from the neighbouring eastern 
and south eastern valley contribute to the smaller morning peak based on analysis 
of the diurnal wind patterns. Gap flows through the western passes of the 
Kathmandu valley which sweep away the valley pollutants towards the eastern 
passes modulated by the mountain – valley wind system are mainly responsible 
for the main pollutant circulation patterns exhibited within the valley.

Meteorological variables at the mid-hill location also showed a similar day – 
to – day pattern with a strong prevailing south-westerly wind during the 
afternoon, and a semi-diurnal variation of surface pressure throughout the 
experiment. The pre-monsoon period was found to be the driest and gustiest 
season. Further studies of pollutants residence time within the valley, its re-
circulation and inter-valley transport timescales will be useful for developing 
future guidance policies for air quality improvement in the Kathmandu valley.
1. Introduction and Background

Urban air pollution is a major concern in the developing world due to its potential effects on air quality, human health and consequences for regional climate change. Such effects are becoming increasingly pronounced in the rapidly developing region of Nepal, and in particular the Kathmandu valley where population growth and urbanization has shown dramatic rises. There is limited published literature available on air pollution characteristics within the Kathmandu valley, but it will be useful to briefly review this in context with current concerns, in particular the potential additional impact due to long range transport of pollutants (Carrico et al., 2003). The mountain – valley wind system and monsoon rainfall are considered a major driver in controlling pollutant circulations and also its dispersion in the region (Nakajima, 1976; Panday and Prinn, 2009; Shrestha et al., 2002; Shrestha et al., 2000; Shrestha et al., 2010).

Effective monitoring of atmospheric aerosols and meteorology at mid and high altitude locations in topographically complex regions such as Nepal provides insight into both regional as well as potential climate impact responses to air pollution and hydrological feedbacks there. A number of studies have highlighted the role of complex topography and associated circulation patterns in the transport of aerosol in mountainous terrain (Beniston, 1987; Gautam et al., 2011; Grigoras et al., 2012; Whiteman, 2000) since aerosol dispersion in such regions are significantly different compared to others (Reid, 1978). In complex terrain aerosol emissions can often be trapped within a valley for sustained periods, localising their properties and impacts and these may be particularly acute in the Nepalese Himalayas (Shrestha et al., 2010), however, explicit understanding requires more in situ observations.

In mountain valley terrain it is generally observed that daytime anabatic winds transport aerosol emissions within the valley to higher altitude whilst katabatic winds re-circulate them during the night time. This mechanism has been well documented in the Sierra-Nevada Mountains and Great Basin, in the USA (Kim and Stockwell, 2007). A similar process can be expected in the Himalayan
foothills of the Kathmandu valley which the study here is a part of. Since local topographic variation can often result in locally unique patterns and interplay between pollutant emissions and transport long term monitoring is required. This is particularly true of the Kathmandu valley where, due to the extremely large altitudinal variation, general theory and findings from elsewhere may not be fully transferrable (Panday, 2006). Hence, in this paper we aim to provide initial information and insight on the seasonal and diurnal characteristics of aerosol number concentration and size distribution in the Kathmandu valley mid-hill region.

It is generally established that increasing aerosol number concentration may suppress precipitation locally and regionally (e.g. Li et al., 2011) as these act as a Cloud Condensation Nuclei (CCN) and Ice Nuclei (IN) which influence properties of cloud and subsequent precipitation development (Lohmann and Feichter, 2005). Due to the heterogeneous nature of aerosol emissions and their dispersion characteristics at the global and local scales (Latha and Badrinath, 2005) this creates major uncertainties in modelling regional climate requiring targeted modelling and observational validation.

Previous aerosol measurements in Nepal have focused on the high altitude regions where local anthropogenic aerosol emissions are almost negligible due to low population density and these experiments have focused on chemical composition and mass concentration. However, aerosols number size distribution, an important parameter e.g. to model aerosol – cloud – precipitation interactions is not well documented across the region. A short term study of surface meteorology and total particle concentration (using a condensation nucleus counter, CN) was carried out by Hindman and Upadhayay (2002), across the Himalayas including the Kathmandu valley and its eastern rim. These showed a semi-diurnal variation with morning and evening maxima and afternoon minima. It was found that the twin peaks were associated with stable air conditions across the valley and minimum concentration was attributed to the valley – mountain wind circulation and strong day time convection. The more detailed results presented here largely confirm this observation.
Previously high concentrations of ultrafine particles (in the size range 10<\(D_p<700\text{nm}\)) has been observed in the high Himalaya region (5079m asl) during a 16 months observation of pollutants at the Nepal Climate Observatory Pyramid (Venzac et al., 2008). They further reported that a maximum diurnal particle concentration was observed during late evening (~2200 LST) and that the pre-monsoon period was the most polluted. A further analysis of two years observation of aerosol concentration (10nm – 10µm) at the same location (Sellegri et al., 2010) confirmed this seasonal variation as did a satellite remote sensing analysis (Shrestha and Barros, 2010). Daily twin peaks of aerosol number concentration (14<\(D_p<340\text{ nm}\)), one in the morning and other in the evening during the pre-monsoon season was observed in the middle mountain region of Nepal (Shrestha et al., 2010). Furthermore, they also reported that the afternoon and night time minima were attributed to the valley – mountain wind circulation. A similar diurnal variation of total PM\(_{10}\) was measured in the Kathmandu valley (Panday and Prinn, 2009). In the study presented here we will add to these observations focusing on the size variation between fine and coarse size ranges in more detail.

In section 2 we will describe the meteorology of the Kathmandu valley as observed during the study period and in longer climatological context. The general characteristics of the experimental site and local valley are explained in section 3, whilst details of the experimental set up are discussed in section 4. Section 5 summarizes the analyses and results followed by discussion and conclusion in section 6.

2. General meteorology of the Nepal Kathmandu Valley

The Kathmandu valley is dominated by a monsoon driven subtropical climate. The Indian summer monsoon generally manifests from June to September, and is responsible for more than 80% of rainfall in the valley (Shrestha, 2000). Southwesterly warm and moist air originating from the Bay of Bengal encounters the complex terrain of the Himalayas triggering this summer monsoon rainfall, however, significant spatial variation in rainfall is observed.
within the valley (Ichiyanagi et al., 2007), as is also found in the complex terrain of the northern Sonara, Mexico (Gebremichael et al., 2007) due to the North American monsoon.

The pre-monsoon season (March – May) is generally dominated by violent lightning and thunderstorms due to intense surface heating subsequent associated convection. In general, the valley receives approximately 1500 mm rain annually and average temperatures are ~18°C (Alford, 1992; Shrestha et al., 1999). The annual temperature range shows a maximum ~35°C during the month May and is below freezing during January mornings. December through February is the winter season and is generally characterized by dry and cool weather with light rain over the valley and snow fall across the surrounding hills and mountains. Winter precipitation is caused by both the winter monsoon and the extra-tropical cyclone locally known as the western disturbance.

During the monsoon season the valley generally experiences cloud free conditions with greater visibility as a result of scavenging of pollutants by the monsoon rain leading to cleaner air within the valley compared to other seasons. A number of studies have also highlighted the important role of rainfall in pollution removal in complex topography (Min et al., 2005; Saha and Moorthy, 2004). In the winter season the valley is relatively cool, dry and dusty, and on most of the days is covered by haze with significantly reduced visibility which can disrupt domestic and international flights. During night time the valley remains calm with clear skies accompanied by a strong temperature inversion that both traps pollutants at the valley floor and triggers dense fog lasting up to 0900 or 1000 LST in the morning. However, this effect is not generally observed in the valley rim. During the pre-monsoon season strong atmospheric circulation brings dust laden air from neighbouring deserts (e.g. from India and Pakistan) and also transports intercontinental dust to the region from the Sahara desert (Carrico et al., 2003).
3. The Kathmandu Valley

The Kathmandu valley is a bowl shaped former lake. It has a surface area of approximately 350 km² with an average altitude of 1,300 m asl. The valley is surrounded by a mountain range with altitudes ranging from 2,000 – 2,800 m asl. As the valley is surrounded by the other smaller, lower valleys this results in complex circulation patterns, but in general, the valley acts as a plateau pulling in air masses during the day time and as a basin during night time, forming a very deep cold pool (Panday, 2006). A location map of the Kathmandu valley, including the neighbouring eastern valleys and the current measurement site, is shown in Figure 3.1. There are number of passes within the mountain range that act as entrances and exits of pollution transportation (Figure 3.2). The valley is dominated by the 3 major cities of Nepal, Bhaktapur, Lalitpur and the capital Kathmandu. Recent census estimates the population to be more than 2 million and rising.

![Location map of the measurement site (Nagarkot, Lat: 27.7°N, Lon: 85.5° E, Alt: 1900m) located in east Kathmandu (KTM) valley. The NCO-P is the Nepal climate observatory pyramid situated at the base of Mt. Everest. Dhulikhel (DKL) and Banepa (BNP) is the immediate neighbouring eastern valley. Elevation contours are plotted as 1000: 500: 6000m.](image-url)
The Kathmandu is becoming a rapidly urbanised Himalayan foothill valley where transport infrastructure is dominated by fossil fuel powered vehicles whose numbers have increased significantly over the last decade. A previous study suggested the number of vehicles increased by more than 60% during the 15 years period from 1989 to 2004 (ADB/ICIMOD, 2006). The local population’s activities generally include preparation of two large meals per day, one in the morning and one in the evening with the majority of home energy sources for cooking being LPG gas, Kerosene and firewood with only a small amount of hydro-electricity capacity. Previous studies within the valley showed that vehicular emissions was the major contributor of local air pollution, followed by household emissions and then the industrial sector, with significant contributions to the latter from the brick kiln factory production (Panday and Prinn, 2009). Brick kiln factories normally operate from December to May, and there are estimated to be more than 500 alone in the Kathmandu valley. Although the factories are scattered across the valley, the majority of them are located in Bhaktapur, the city located close to the measurements site and will be shown to contribute to ambient air loadings under specific conditions.

**Figure 3.2**: Google Earth image of the Kathmandu Valley viewed from the south showing the main peaks and passes. Nagarkot peak is the location of the measurement site. The acronym Ktm, Ptn and Bhk refer to the three major cities in the Valley Kathmandu, Patan and Bhaktapur respectively.
4. Experimental setup

The measurement site – Nagarkot is located in the eastern flank of the Kathmandu valley (Lat 27.7° N, Long. 85.5° E). The altitude of the measurement site is about 1,900m asl and is located in the so-called “middle hill” region of central Nepal, approximately half-way between the Indo-Gangetic plain and the Tibetan plateau. At this site a small monitoring station consisting of an optical particle counter (OPC) (GRIMM Model 1.109) with a PM-10 inlet, and a Vaisala meteorological sensor (Met Pack WTX 510) were installed on the roof top of a two storey public health office building. The station was located approximately 5m above the ground. The station was mounted mast above the corner of the rooftop parapet offering good 360° exposure. Data was recorded to a notebook computer for the period March to December 2011.

Aerosol particle concentrations in 31 different size channels ranging from 0.25µm to 32µm. The GRIMM OPC uses a light scattering technique to measure and count individual particles. The air sample is drawn in via a flow-controlled pump at a rate of 1.2 l/min. This is then passed through a laser beam where the particle scattering signal is classified using a pulse height analyzer and stored in appropriate size channels. Particles size distributions were recorded at 1 Hz and averaged over 1 minute. In addition the OPC station provided additional measurements of ambient relative humidity and temperature again with a temporal resolution of one minute.

The weather station provided measurements of temperature, relative humidity, surface pressure, wind speed using an ultrasonic anemometer, wind direction, and a precipitation sensor with rainfall and hail discrimination using piezoelectric sensor which measures electrical charge induced from pressure exerted by individual drops which is proportional to the volume of the drops. As with the OPC the meteorological station was set to record variables at 1 minute intervals. Data presented here will be based on median values calculated based on these one minute averaged data measured by the OPC and Vaisala weather sensor unless it is specified.
5. Results

5.1 Local Meteorology

5.1.1 Seasonal Variation

Figure 3.3(a-c) and Figure 3.4(a-c) showed the time series analysis of meteorological variables measured from March through December 2011. A statistical summary is provided in Table 3.1. Pressures are at the local altitude to record variation, and are not corrected to sea level. Winds were generally strongest during the pre-monsoon season. Maximum gusts occurred in April and could reach greater than 20 m/s. These high gust periods were attributed to the strong thermal circulation in the valley associated with high daytime surface temperatures followed by strong night time cooling. In general the prevailing wind direction was from west to southwest. Very little rainfall was recorded in the beginning of the season and any events were of low intensity however towards the end of the season rainfall intensity increased significantly reaching as high as 130 mm/h. A considerable amount of haze was also recorded during the middle of the period and through to the end of this season which is a general feature of the region prior to onset of the monsoon. Atmospheric pressures fluctuated between 801hpa to 815hpa and surface temperatures generally increased over time with maximum reported temperatures in March ∼19°C which increased to 25°C by the end of May. Similarly relative humidity (RH) was highly variable in the beginning of the season but on average increased significantly towards the end of the season indicative of the arrival of the monsoon season.

With the arrival of the monsoon season winds decreased in strength, with means ∼1.9 m/s however maximum gusts up to 25 m/s were observed at the beginning of the season after which wind speed decreased with calm wind conditions often occurring by the end of the season. The data showed that wind was coming from all directions particularly in middle of the season. During the beginning and end of the season NE – SW and SW – NW wind directions were
dominant. Rainfall was generally mild but continuous, again a typical characteristic of the Indian monsoon, and no hail events were recorded. Atmospheric pressure was lower than the other seasons, ~805 hpa, with more stable pressure systems compared to the other seasons. Temperature also displayed a stable trend throughout the season with maxima close to 25°C, similar to the pre-monsoon season. In general RH remained high throughout the season, above 80% in most of the cases.

Figure 3.3: Time series analysis of meteorological variables a) Wind speed (m/s), b) Rainfall (mm/hr), and c) Wind direction (deg). The gap is due to power outages.

At the beginning of the post monsoon season winds gradually moved NW and NE for a short period, however, during the end of the season SW winds became established. A sharp increase in wind speed from the beginning of the season occurred and increased gradually over time. Mean wind speeds were 2.9 m/s with maximum gusts of 15.9 m/s. Several episodes of low intensity rain and hail were also recorded. A sudden rise in the pressure was observed from the
beginning of the season which remained stable over the entire season. The highest recorded pressure over the year was recorded during this period. Temperature showed a decreasing trend over time with maximum values around 22°C at the beginning of the season, however this dropped to almost 8°C by the end of November. The RH decreased compared to the monsoon season with the most pronounced decrease, ~20%, occurring towards the end of the season.

In the winter season NE – NW winds with mean speeds ~3.4 m/s, were dominant, with the strongest winds observed over the year, with maximum gusts ~14.4 m/s. No rainfall was recorded during this season at this altitude. Atmospheric pressures dropped relative to the post-monsoon season fluctuating between 806 – 815.5 hpa. Surface temperatures dropped steadily from the post-monsoon values until the beginning of the winter before gradually increasing. RH levels generally decreased during the season, dropping by as much as 14% on average.

Figure 3.4: Same as Figure 3.3 but for a) Relative Humidity (%), b) Surface Temperature (°C), and c) Atmospheric Pressure (hpa), local altitude value.
5.1.2 Diurnal Variation

Figure 3.5(a-d) and Figure 3.6(a-d) show the diurnal cycles of the mid-hill wind during the different seasons. We observed that diurnal patterns of wind are dominated by mountain–valley circulation as the wind speed increases with increase in intensity of solar heating. During the pre-monsoon season (March–May) it was found that during the early morning light winds (<10 km/hr ~2.77 m/s) blew from the NE–SE originating from the surrounding hills as a continuation of the general katabatic flows. During the day Nagarkot begins to be influenced by upslope flow and wind speeds gradually increase reaching maximum strength during the afternoon, typically ~4 m/s in March but with gusts occasionally greater than 20 m/s.

We observed that up slope flow starts earlier in the pre-monsoon than in the other seasons. It was found that the morning fog associated with the temperature inversion dissipated earlier in the day than in the winter season which may contribute to generation of the early onset of upslope flows. The measurements showed that the SW wind from morning to dusk, which sometimes persisted until midnight in some months, was associated with upslope and up-valley winds flowing through the Bagmati River gorge, which is located in the SW of the valley, and by gap winds through the western passes of the valley. Previous research (Panday and Prinn, 2009; Regmi et al., 2002) also observed these winds from these passes in the late morning.

During the monsoon season (June–Sept.) we observed significant daily variability in wind direction. Figure 3.6(a-d) shows the wind direction quartile range is significantly higher in the monsoon period than in any other season and also reveals in the Figure 3.7(a). Wind were generally calm to light throughout the day but still dominated by thermal circulation and displayed a clear diurnal cycle with light winds in the peaking at ~4 m/s in the afternoon. The maximum gusts in this period could exceed 25 m/s during the afternoon at the beginning of the season consistent with the transition from the end of the pre-monsoon to early monsoon period which is characterized by a stronger atmospheric circulation and generation of severe thunderstorm and lightning activity.
Figure 3.5: Wind speed in m/s a) Pre-monsoon, b) Monsoon, c) Post-monsoon, and d) Winter.
In the post-monsoon season (Oct. – Nov.) NE winds were found to occur early in the morning that quickly backed towards the SE and then stabilising as they strengthened into SW winds in the middle of the morning remaining established until late evening. The winds in general were found to be stronger than in the monsoon season but still strongly dominated by thermal circulation. During the afternoon as shown by the 75\(^{th}\) percentile analysis, wind reached close to 4 m/s, followed by a sharp decrease over time until late evening and again gradually increased until the next morning finally peaking ~ 3 m/s in the early morning time.

During the winter season (Dec. – Feb.) easterly winds were dominant until the middle of the morning. Unlike the other seasons however winter wind patterns were very different (Figure 3.7(b)). Figure 3.6(d) shows that a SW flow begins later in the morning, attributed to the reduced solar heating at the surface during winter, delaying the onset of upslope flow. Moreover, delayed initiation of the upslope flow was also caused by the prolonged winter morning fog associated with the strong cold air pool and temperature inversion that formed in the valley. The SW wind remains for only a short period of time, moves to SE in the early evening before veering to a downslope easterly wind, suggested that the warming of the valley atmosphere plays a significant role in perturbing the circulation at this time of year. In contrast to the other seasons, where strong winds occurred during the afternoon, winds were found to be strongest during the morning, typically ~4 m/s, whilst slightly weaker wind, ~3 m/s, were experienced during late afternoon. Strong gust were also recorded during the early morning period, exceeding ~14 m/s.

The atmospheric pressure measurements (not shown) displayed a clear diurnal cycle throughout the year. We observed lowest atmospheric pressure in the monsoon season followed by pre-monsoon, winter and post-monsoon in increasing order. Daily twin high pressure peaks occurred at 0900 and 2100 LST and low pressure systems developed at 1500 LST and 0300 LST in almost all seasons. Analysis showed that the diurnal fluctuation was more pronounced than the nocturnal variation. The maximum daytime fluctuation was found to be ~3.5 hpa during the pre-monsoon season. The night time pressure was relatively stable.
as compared to the day time and was found to be \(~1.6\) hpa during the monsoon season.

**Figure 3.6**: Diurnal cycle of wind direction for a) Pre-monsoon, b) Monsoon, c) Post-monsoon, and d) Winter seasons.
In the pre-monsoon season morning peak pressure was ~811 hpa which generally decreases over time and reached an afternoon low of 807 hpa. The evening peak pressure is found to be lower than the morning peak and night-time low is slightly higher than the afternoon low. During the monsoon season the morning high pressure could reach up to 805 hpa and the afternoon low was found to be around 802.5 hpa. The evening high was observed nearly the same as the morning high whereas nocturnal low pressure was recorded higher than the afternoon low. In contrast to the other seasons where day-time maxima were found to be noticeably larger than the night-time maxima, and day-time minima were smaller than night-time minima, during the monsoon season day-time high pressure was found to be equal or even slightly lower than the night-time high pressure.

In the post-monsoon period day-time high pressure was observed to be the highest over the year, > 813 hpa, whereas daytime low pressure was measured around 810 hpa. In the winter season the morning pressure was found to be around 812 hpa which gradually decreases over time and reached to day time minima with magnitude of 809 hpa. The night time high pressure was recorded to be lower than the morning high and low pressure was found to be slightly higher than the day time minima.

The temperature variation at the mid-hill station site showed clear diurnal cycles and as well as a distinct seasonal pattern. In general maximum temperatures were recorded during the middle of the day as expected whereas lowest temperature was during early morning at around 0500 LST. The maximum diurnal fluctuation in temperature was found to be 7.3°C in March whereas minimum temperature variation was 3.6°C in July. In the pre-monsoon season maximum temperature was recorded around 20.5°C in May afternoon which decreases sharply until evening. The minimum temperature recorded was 11.25°C in early morning in March.

In the monsoon season the maximum afternoon temperature recorded was 21.7°C in June whereas the minimum morning temperature was 16.6°C in September. The mean temperature gradually decreased over time with the post-
monsoon afternoon maximum of ~19.6°C occurring in October afternoon whereas the minimum temperature was ~10°C occurring in the November. In the winter season the daytime maximum temperature was 13.1°C in December whereas early morning minimum temperature was 6.6°C.

The diurnal cycle of relative humidity (RH) revealed (not shown) a minimum RH in the afternoon and maximum in the early morning. We observed the maximum diurnal cycle of RH occurred in March, more than 33%, whereas the minimum cycle, 10%, occurred in July. Furthermore, it was also observed that March was the driest month and July the wettest month on average at this altitude over the year based on these recordings.

**Figure 3.7**: Wind rose a) Pre-monsoon, and b) Winter season.

In the pre-monsoon season the average minimum RH was ~ 30% in March with maxims in May (during the morning) ~ 82.5%. During the monsoon season RH generally increases throughout the day. Minimum RH was ~ 70% in June whereas the maximum RH was observed in July, ~ 92%. Most of the time RH exceeded 85% particularly from early evening to morning throughout the monsoon season. The RH then begins to decline from the beginning of post-monsoon season. In general this season remained relatively dry during the day
with clear nights. A maximum and minimum RH was recorded, 88% and 59%, respectively during October, however, the majority of the data shows RH remaining at >80% from early evening to next morning. In the winter season the RH could drop by up to 46% in the afternoon and exceed 70% in the early morning.

5.2 Aerosol Characterization

5.2.1 Mechanisms Influencing Dispersal in the Kathmandu Valley

We observed several possible mechanisms operating over the day to circulate and/or disperse pollutants in and out of the Kathmandu valley. It was found that when the temperature inversion was broken up during the morning period surface pollutants over the valley as a whole increase and is characterised by a “brown haze” which is diluted over time due to the increased mixing as the boundary layer height grows. This was also observed by Panday (2006). We were also able to capture some of these possible mechanisms with images taken during early period of the field experiment which is characterized as a dry and polluted season. Here we will discuss several mechanisms influencing the pollution build-up and dispersal before examining the supporting aerosol measurements.

Figure 3.8: Photograph taken on 28th of February 2011 in the late morning from the Nagarkot station viewing towards the west of the Kathmandu valley. The “brown haze” layer fully covered the Nagarjun peak (2,100 m asl).
Firstly as the valley experiences increased heating by solar radiation a strong convective current is formed coupled with an increasing mixing layer height. This process generally occurs over the flat topography, an initially within the valley the same process begins with initially little influence from on up and down slope flows. The mixing layer can eventually grow higher than the valley side wall transporting pollutants out of the valley which in turn are carried away by the upper level winds. A similar thermally driven mechanism was observed across the UK (Dacre et al., 2007) and the Amazon rain forest (Edy et al., 1996). The Kathmandu valley, which has a ratio of vertical relief to horizontal expanse of ~ 1:30, can serve as a flat plain to allow this process. Figure 3.8 shows the growth of the mixing layer height to altitudes greater than the surrounding mountains transporting pollutants higher up and over the valley floor. The photograph was taken in the late morning viewing west from the Nagarkot. The Nagarjun peak (2,100m asl) was completely obscured by a brown haze indicating that pollutants had been transported up to 1km above the valley floor. Previous observations and modelling studies in the valley also suggested that the mixing height could reach up to 800m from the valley floor (Panday et al., 2009; Regmi et al., 2002).

The second mechanism that can contribute is a similar process, but here the mixing layer instead of overtopping the valley side wall is confined below the main peaks. In this case the pollutants cannot be efficiently ventilated from the valley. Instead they become trapped and re-circulate within the valley complex over time. This effect is observed in the North Saskatchewan River valley in Edmonton, Canada, where pollutants recirculate in the evening resulting in pronounced evening maximum concentrations (Rudolph and Hage, 1983). This process is hypothesized to be responsible for an evening pollution peak because recirculation of aged pollutants combine with the fresh evening emissions. Figure 3.9 shows a case where the “brown haze” layer remained below the height of surrounding mountains.
The third mechanism that can contribute to pollution levels is from gap flow through the western passes of the Kathmandu valley. The valley consists of a number of passes in the western and eastern rims that bring strong westerly winds into the valley. Westerly winds from the Bhimdhunga and Nagdhunga passes can contribute to dispersal of the valley pollutants via the eastern passes e.g. the Nala and Sanga passes. Our analysis also showed that SW winds from late morning to evening are dominant almost throughout the year consistent with the direction in which these passes lie in relation to the station. Figure 3.10 shows that the western part of the valley is free from haze whereas the eastern part still retains pollutants. This photograph was taken in the early afternoon when the strong westerly wind had fully developed by that time.
Figure 3.10: Photograph taken on 8th March 2011 in the early afternoon. The western passes of Kathmandu valley Bhimdhunga and Nagdhunga are relatively haze free compared to the SW of the valley indicating the westerly wind has swept pollutants through the eastern passes to the neighbouring Dhulikhel and Banepa valleys. The pollutants are completely swept dispersed by the middle of afternoon (Figure 3.11).

Figure 3.11: Photograph taken on 8th of March 2011 during middle of the afternoon showing the aftermath of westerly winds dispersing the pollution.

A fourth mechanism that can contribute to pollutant removal is via strong upslope and up-valley flows that develop during the day-time. The pollutants are transported higher up towards the mountain peaks. A number of studies have highlighted the role of these thermally driven upslope winds in pollutants transport. A numerical study of pollution recirculation processes in the lower
Fraser valley in British Columbia, Canada, demonstrated that significant pollution can be transported vertically by upslope flows although in that case the pollutants were re-circulated back to the valley surface (Borrego and Norman, 2007).

Without a similar detailed modelling analysis it is difficult to predict the dominant mechanisms responsible for removing pollutants from the Kathmandu valley. However, it seems from this brief analysis that multiple mechanisms are acting and can be responsible for pollutant circulation into and out of the valley. We will try to further interpret the presence of these mechanisms in section 6 with the help of the aerosol measurements.

5.2.2 Seasonal Variation

Figure 3.12(a-c) shows the time series analysis of aerosol concentration measured during the experiment from March till December. Table 3.2 provides the statistical summary of the measurements. During the pre-monsoon season the coarse particle concentration (D_p>2.5µm) was found to be less than 50 cm⁻³ in most cases but could reach up to 800 cm⁻³ in some extreme cases. The contributing concentration for fine particles (0.25 <D_p< 2.5 µm) was found to be dominant in this season. We observed that this fine particle contribution to the total could exceed than 6,500 cm⁻³ in some cases. Data from mid-November to December are omitted from the analysis due to an instrument failure.

Figure 3.12(a) showed that during the monsoon season the frequency and concentration of coarse particles increased substantially whereas the fine particle contribution showed a slight decreasing trend over time. Mean total particle concentrations in this season were found to be lower than other seasons, suggesting that the monsoon rain plays a significant role in scavenging pollutants as expected. This effect was previously observed in central Nepal by Shrestha et al. (2010). Maximum total particle concentrations exceeded 8,000 cm⁻³ in some extreme cases, but this was rare. It was also observed that mean coarse particle concentrations were higher than in the other seasons whereas mean fine particle concentrations were lowest in this season.
Figure 3.12: Time series analysis of pollutants a) Showing fine particle concentration, \((0.25 \leq D_p \leq 2.5 \mu m)\), and coarse particle concentration \((D_p>2.5 \mu m)\) \((cm^{-3})\), b) Aerosol number concentration \((cm^{-3})\) as a function of particle size \((\mu m)\), and c) Total aerosol concentration \((cm^{-3})\).

During the post-monsoon season it was observed that the total particle concentration generally began to increase but also revealed an interesting sub-seasonal monthly cycle. For example both coarse and fine particle concentrations reached their peak in the mid-October and declined during the end of that month and so on. The contribution of coarse particles was not considerably high in this season. It was found that the post-monsoon season was the most polluted with mean particle concentrations \(> 900 cm^{-3}\) and which were dominated by fine particles. Contribution of the coarse particles was very low with observed maximum concentrations \(\sim 300 cm^{-3}\).
5.2.3 Diurnal Variation

Figure 3.13(a-c) showed that in the pre-monsoon season total particle concentrations exhibited distinct diurnal twin peaks, one in the morning (0500 – 0800 LST) and the other in the evening (1800 – 2200 LST). The evening peak was generally more pronounced than the morning peak. Furthermore, it was observed that the nocturnal concentration was higher than the diurnal pollutant concentration. The maximum concentration was recorded in March (evening period) which could reach 1,100 cm$^{-3}$ whereas during the end of the season it had consistently dropped to around 800 cm$^{-3}$, suggesting a significant intra-seasonal variability of pollutants concentration in the Kathmandu Valley arising from one or more of the mechanisms discussed above. The morning peak concentration was typically $\sim$ 900 cm$^{-3}$ but this decreased during the end of the season. Discounting occasional enhanced events concentrations were observed to be $\sim$500 cm$^{-3}$ and this appears to represent a typical background level for aerosol pollutants during the pre-monsoon season at this altitude.

We also analyzed the variation in particle size distributions recorded in the morning and evening peak periods (Figure 3.16(a-c)). Although the fine and accumulation mode particles (0.1 – 1µm) are dominant in the both, the contributions of coarse mode particles are largest during the morning peak than in the evening peak. We compared the morning and evening peak particle size distribution to the diurnal and nocturnal minimum distribution and it was generally found that enhancements in all sizes of particles have contributed to both peaks.

PM$_{10}$ mass loadings were also estimated (assuming perfectly spherical particles with average density of 1.5 gcm$^{-3}$, Figure not shown). This again showed that the PM$_{10}$ followed a similar diurnal cycle as the total particle concentration with daily twin peaks in the morning and other in the evening. A seasonal evening maximum PM$_{10}$ with a magnitude of $\sim$68.5 µgm$^{-3}$ was found for March. In contrast the morning PM$_{10}$ peak was found in May, 46µgm$^{-3}$. Outside these peaks typically greater than 30 µgm$^{-3}$ was representative of background loadings.
throughout the season. Volume and area distributions showed that diurnal peak (non-peak) of the pollutants are mainly driven by the presence (absence) of coarse particles (Figure 3.14(a-c)) in this season.

Figure 3.13: Diurnal variation of total particle concentrations observed during the a) Pre-monsoon, b) Monsoon and c) Post-monsoon seasons.
Figure 3.14: Volume distribution of the pollutants represented by median (50th Percentile shown in dark colour), upper quartile (75th Percentile shown in light grey colour) and lower quartile (25th Percentile shown in light grey colour) a) Pre-monsoon, b) Monsoon, and c) Post-monsoon season.
The pollution rose analysis of total particle concentration (cm$^{-3}$) is shown in Figure 3.15(a-c). Aerosol levels recorded for 210 – 270°N, SW-W wind directions, were mainly responsible for transporting polluted urban air mass from the city center. As mentioned previously, the rose diagram also shows background concentrations exceed 500 cm$^{-3}$ in the pre-monsoon period.

**Figure 3.15**: Pollution rose a) Pre-monsoon, b) Monsoon, and c) Post-monsoon season. Error bars are represented by one Standard Deviation.
In the monsoon season the diurnal variation in total particle concentration showed a similar pattern as in the pre-monsoon season. We observed evening and morning peaks occurred during 1700 – 2100 LST and 0600 – 0700 LST respectively, almost the same as in the previous season. In contrast to the dominant evening peak in the pre-monsoon season both peaks are now almost similar in magnitude or with a slightly higher morning peak. We also observed maximum nocturnal rainfall occurring during this season. The morning peak concentration during the monsoon onset period was found to be more than 575 cm\(^{-3}\), which decreased over time to less than 360 cm\(^{-3}\) during the end of the season. Similarly the evening peak concentration was 580 cm\(^{-3}\) during the monsoon onset period again dropping to 330 cm\(^{-3}\) by the end of the monsoon season. The minimum particle concentration during the monsoon season was 175 cm\(^{-3}\) indicating a very low background particle concentration during this season.

The monsoon season was characterised by a greater contribution by coarse particles as shown in the time series diagram in Figure 3.12(a). Particle size distribution analysis during the daytime peaks and the nocturnal periods showed that concentrations of these coarse particles were significantly increased. Analysis of the slope of the size distributions showed it to be greatly reduced and the volume distribution significantly increased as compared to the other seasons (Figure 3.16(a-c)). However, it was also observed that the coarse particle contribution was substantially reduced during the afternoon periods. The daily twin peaks and nocturnal concentrations at the Nagarkot site are attributed to the transport of fresh aerosol emissions from biomass burning and Brick factory activity in or near the city center, which were also impacted by scavenging by the monsoon rain. The volume and area distributions also decreased during the afternoon relative to the morning and nocturnal periods.
Figure 3.16: Size distribution of aerosols showing the contribution of different sizes of particles for the peaks and minimum and maximum particle concentration periods during the a) Pre-monsoon, b) Monsoon and c) Post-monsoon season.

We observed a slightly different signature for PM$_{10}$ during the monsoon season, however it still showed a clear diurnal variation with a morning and an evening peak with low concentrations from morning to early afternoon which gradually increased over time, reached a peak in the evening and again began to decline. Nocturnal higher background concentrations were attributed to the
presence of coarse particles during the night time. The diurnal minimum PM$_{10}$ was found to be less than 15 $\mu$g m$^{-3}$ in September whereas night time minimum concentrations were always greater than 60 $\mu$g m$^{-3}$. A maximum seasonal PM$_{10}$ evening peak was found in July which was calculated to be more than 3,000 $\mu$g m$^{-3}$. We also observed that the 75th percentile of the data shows several orders of magnitude more than the median PM$_{10}$ concentration, during the evening to early morning time. The maximum seasonal morning peak was found to be more than 900 $\mu$g m$^{-3}$ in July.

The concentration wind rose diagram showed significant intra-seasonal variability. We found high particle concentrations in September whereas low concentration was observed in July. However, overall we observed low mean particle concentrations in the monsoon period compared with other seasons, as reported in the Table 3.2. In general SW – NW winds were responsible for transport of aerosol pollutants to the site.

In the post-monsoon season aerosol concentration showed quite similar patterns as in the monsoon season. We found similar daily twin peaks, one in the morning and the other in the evening as in the previous season. Unlike the monsoon season, where morning and evening peak concentrations were quite similar, the evening peak concentration was more pronounced than the morning peak. It was found that the evening peak was typically > 900 cm$^{-3}$ whereas morning in the peak was observed to be slightly less on average, 775 cm$^{-3}$. Moreover, the nocturnal minimum concentration (625 cm$^{-3}$) was found to be almost double the diurnal minimum concentration (330 cm$^{-3}$), indicating increased background pollution concentration compared with the monsoon season (175 cm$^{-3}$).

The PM$_{10}$ concentration (not shown) followed a similar pattern as the total particle concentration with a morning and an evening peak of, 43 $\mu$g m$^{-3}$ and 49 $\mu$g m$^{-3}$ respectively. The night time minimum concentration was found to be typically greater than the day time minimum concentration.
Referring to the particle size distributions during peak and minimum concentration time, in general, the peak concentrations were attributed to the increase in all size particles both in the morning and in the evening. However, we observed significant increases in coarse particle concentration during the evening and at night.

The aerosol concentration rose diagram showed that background concentrations were increased compared to the monsoon season, and were typically around 500 cm$^{-3}$. The majority of these were transported to the site by SW – NW winds.

6. Discussion and conclusions

From March to December 2011, aerosols number size distributions (0.25μm – 32μm), wind speed and direction, temperature, relative humidity, atmospheric pressure and rainfall were continuously measured at Nagarkot (Lat: 27.7°N, Lon: 85.5°E, Alt.: 1900 m), Nepal, located in the eastern flank of a bowl shaped Kathmandu valley. Our results showed a distinct seasonal and diurnal cycle of aerosol pollutants in the valley. We observed daily twin peaks throughout the year, one in the morning and other in the evening. The maximum aerosol number concentration, dominated by the fine particles ($0.25 \mu m \leq D_p \leq 2.5 \mu m$), occurs during the dry season whereas seasonal minimum particle concentration with a significant contribution of coarse particles ($D_p > 2.5 \mu m$) occurs during the monsoon season. In general it was found that the fine particles are dominant in all seasons. Our analysis showed that the post-monsoon season ($919 \pm 694 \text{ cm}^{-3}$) was the most polluted followed by, in decreasing order, the pre-monsoon ($746 \pm 569 \text{ cm}^{-3}$) and monsoon season ($379 \pm 361 \text{ cm}^{-3}$). The analysis shows that the Kathmandu valley receives significant PM$_{10}$ input during the monsoon season (typically > 30 μg m$^{-3}$) which is contributed to by coarse particles from local sources. We also observed significant intra-seasonal variability in the aerosol concentrations.
The diurnal peaks in aerosol concentration are attributed to morning calm wind conditions followed by the transitional growth and break down of the nocturnal boundary layer. We found that both peaks are influenced by enhancement of the coarser particles. The evening peak, generally higher than the morning peak, and is caused by the light SW winds blowing through the Kathmandu valley carrying fresh evening emission from increased local activities and suppression of the mixing layer, which recirculates aged pollutants trapped. A number of previous studies in the Kathmandu valley also suggested that the peaks are caused by boundary layer processes (Shrestha et al., 2010) and pollutant recirculation (Panday and Prinn, 2009). This study at higher altitude supports this with the morning peak influenced by the growth of the mixing layer. Furthermore, air mass transported by up-valley wind from the neighbouring eastern and south eastern valleys, as suggested by the diurnal wind patterns, contributes to the morning pollution peak. The low concentrations observed during the afternoon are attributed to the strong south – westerly wind that sweeps away the valley’s pollutants towards the eastern passes. However, nocturnal high aerosol concentrations are caused by a combination of a stable atmosphere enhanced by shallow boundary layer height. We suggest that the third and fourth mechanisms for dispersal, described in the section 5.2.1, are mainly responsible for pollutant circulation in the Kathmandu valley.

Meteorological variables also show a similar day – to – day variability with a strong wind blowing from south to west direction. The SW wind is found to be dominant from morning until dusk with its maximum amplitude in the afternoon. Diurnal variation of atmospheric pressure is caused by the atmospheric tides, showed a similar diurnal pattern throughout the year with peak pressure at 0900 and 2100 LST and minimum pressure at 1500 and 0300 LST. Wind is generally found to be strong during the winter and calm during the monsoon season. The pre-monsoon season is observed as a driest season as expected and also found to be gusty characterized by a violent atmospheric circulation causing heavy downpour. Applying the Kolmogorov-Smirnov (K-S) test for 2 samples to pre-monsoon and post-monsoon seasons we are able to reject the null hypothesis at the 20% significance level and state that the winds are distributed differently at different hours (morning, afternoon, evening and night) of the seasons. Likewise,
the K-S test also confirmed that the total aerosol particle concentrations at the different hours of the seasons are significant at 20% level.

While this study represents a limited set of field measurements, with conclusions being drawn based on only a single station, it is hoped the results may contribute to understanding the local air pollution in the Kathmandu. A detail investigation of aerosol chemical composition would add significantly to the interpretation of local sources and residence times of these within the valley and to assess contributions to and from its inter-valley transport. This would assist in developing abatement strategies for this rapidly developing region.

**Acknowledgement**

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**Reference**


Table 3.1: Statistics of meteorological variables at the Nagar kot, Nepal

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Table 3.2: Statistics of pollutants at Nagarkot, Nepal

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CHAPTER 4: Sensitivity of WRF cloud microphysics to simulate a convective storm in the region of complex terrain

The following chapter will be submitted for publication in Annales Geophysicae.

Authors: Rudra K. Shrestha\textsuperscript{1,2}, Paul J. Connolly\textsuperscript{1}, Martin W. Gallagher\textsuperscript{1}

\textsuperscript{1}Center for Atmospheric Science, School of Earth, Atmospheric and Environmental Sciences (SEAES), The University of Manchester
\textsuperscript{2}Sustainable Consumption Institute (SCI), The University of Manchester, UK

Contribution from others: Paul Connolly helped with the initial model set up and data analysis. Martin Gallagher reviewed the paper and gave critical comments. All work was carried out under the supervision of Martin Gallagher and Paul Connolly.

Abstract

This paper investigates sensitivity of Bulk Microphysical Parameterizations (BMPs) schemes to simulate a convective storm that generally evolves during pre-monsoon season over the complex Himalayan terrain of central Nepal. The Weather Research and Forecasting (WRF) model with explicit convection resolution grid spacing (3km x 3km) is used in this study. We aim to identify an appropriate microphysical scheme to simulate Indian monsoon precipitation from four mixed-phase BMPs schemes (Lin, WSM6, WDM6 and Morrison) which make different assumptions to represent cloud processes in the model and are generally parameterized with an increasing complexity from single to double moments of particles distribution. The simulations were tested for a convective event that occurred in the late afternoon of 18\textsuperscript{th} May 2011. We compared the model outputs with the limited in-situ observations at Nagarkot,
Nepal (Lat: 27.7°N, Lon: 85.5° E, Alt: 1900m). A moderate sensitivity of rainfall to the chosen BMPs was observed, which mainly precipitate out in the windward slope of the hills and confined across the ‘Terai’ (low land in South of Nepal, altitude < 200m), Siwalik and Mahabharat ranges (geographical regions having 10 – 50 km wide swath with altitude ranging from 200 – 3000m extending parallel to south of the Himalayas). Upper level condensate and cloud fraction showed a strong sensitivity to the BMPs schemes whereas relative humidity, surface temperature, wind speed and direction indicated moderate sensitivity. Overall, the sophisticated double moment (Morrison) scheme outperformed the simplified schemes. However, the model performances suggested that more observations are needed to improve our understanding of the microphysical processes in the complex Mountainous terrain.
1. Introduction

One of the major uncertainties in the numerical weather prediction (NWP) is arises from inadequate treatment of cloud microphysical processes in the model (Otkin and Greenwald, 2008; Wevergerg et al., 2010). Although high resolution simulations are made possible by advanced computing power, accurate forecasting can not be guaranteed, became a primary concern while applying the NWP model in research and forecasting of weather. This task is even more challenging over the Himalayas due to role of the complex terrain, where general theory and findings from elsewhere may not be fully transferrable (Panday, 2006). Furthermore, cloud microphysics may enhance storm development through the latent heat release during the process (Andreae et al., 2004; Garabowski et al., 1999; Khain et al., 2005), which is not fully understood yet. Such effects are significant in the cold cloud microphysical processes, which are considered as a dominant precipitation formation mechanisms in the Mountainous region (Chen and Lamb, 1999).

Convective storms are characterized by strong wind and intense rain and/or hail precipitation, which can be evolved as a Mesoscale Convective System (MCS) that could damage thousands of infrastructures, several hectares of agricultural farm leading to reduced crop yield and pose serious threats to human beings. It is estimated that the hurricane Katrina killed more than 1,000 and directly affected ~700,000 people in the United States in 2005 (Gabe et al., 2005). An accurate prediction of MCS, which is more complex than an individual convective storm (Houze, 2004), not only saves lives and billions of properties but also important for disaster prevention plan such as deployment and operation of early warning systems. However, limited understanding of cloud microphysics and its interactions with dynamics in the current numerical modeling practices cause several uncertainties in the weather predictions.

In recent years sophisticated microphysical parameterization schemes have been developed. In the bin microphysics scheme evolution of particle size distribution (PSD) is explicitly resolved (Feingold et al., 1994). However in the
bulk microphysics PSD is represented by a function derived from in-situ measurements (Lin et al., 1983; Rutledge and Hobbs, 1983). Moments of particle size distributions are used to estimate bulk parameters which predicts either mixing ratio (single moment) or mixing ratio and number concentration (double moment) for each class of hydrometeors. Due to high computational advantages over the bin microphysics scheme, the bulk scheme is widely used in the cloud resolving simulations (Morrison et al., 2005).

A wide variety of microphysical parameterization schemes are used in the cloud resolving simulations (Hong et al., 2009; Liu and Moncrieff, 2007; Liu et al., 2011; Rajeevan et al., 2010). The degree of variations in the BMPs schemes are associated with the complexity of microphysical processes that estimates the properties of hydrometeor species (McCumber et al., 1991). So sensitivity analysis of hydrometeor particles to microphysical processes is an important issue which helps to improve current microphysical parameterizations schemes and also guide for further development of the processes.

A number of studies have highlighted sensitivity of cloud microphysics in real data simulation using meso-scale numerical models. A sensitivity analysis of summer-time convection to bulk microphysical parameterizations was carried out using four microphysical schemes coupled with a cloud-resolving meso-scale model covering the continental United States (Liu and Moncrieff, 2007). The simulations were conducted for a week, where microphysics schemes were chosen in such a way that to include simple ice (ice and liquid drop does not coexist) to mixed-phase clouds. The authors emphasized that their simulations showed a greater sensitivity of microphysical parameterizations to ice phase processes.

Reisner et al. (1998) evaluated single and double moment microphysical parameterization schemes using a cloud-resolving (MM5 – Fifth generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model) model to predict super-cooled liquid water during two winter storms across the Rocky Mountain and surrounding areas of Colorado. Comparison of simulated results showed that the sophisticated double moment microphysics with three-class ice scheme was able to reproduce the observed
patterns in the both storms. A comparison of single and double moment bulk microphysics parameterization schemes with five prognostic hydrometeor species (cloud water, rain water, cloud ice, snow and graupel) in the simulations of an idealized stratiform precipitation suggested a strong sensitivity of hydrometeors to the chosen cloud microphysical schemes, where the double moment produced precipitation spread over a wide geographical area than in the single moment scheme (Morrison and Thompson, 2009). A strong sensitivity of wintertime precipitation to the selected cloud microphysical parameterization schemes was also reported across the complex terrain of Colorado Headwater region (Liu et al., 2011).

In the present work we focus on microphysical dependence of meteorological variables in a convective storm that developed over the complex terrain of the Himalayas. This phenomenon has been well documented over continental United States (Liu and Moncrieff, 2007; Liu et al., 2011), North Atlantic Ocean (Otkin and Greenwald, 2008) and Southeast India (Rajeevan et al., 2010) however not been investigated yet in the Nepal Himalayas. Although sophisticated cloud microphysical parameterization schemes are developed in recent times, realism of prediction needs to be tested and verified, particularly this is more important in the Mountainous terrain. Hence, the objective of this study is to examine the high resolution WRF simulation in representing a convective storm that evolved over central Nepal. Herein we evaluate sensitivity of the meteorological variables to the four BMPs available within the WRF model and compare the model results with the limited in-situ measurements.

This paper is organized as follows. Section 2 describes experimental design and also highlights the bulk microphysical parameterization schemes to be tested in this experiment. The case study, including modeling results and observations, is described in section 3. Summary of our findings are presented in section 4.
2. **Experimental design**

A numerical experiment was conducted using the Weather Research and Forecasting (WRF) model version 3.1.1 with Advanced Research WRF (WRF-ARW) dynamical core (Skamarock et al., 2008). The WRF is a non-hydrostatic, primitive equation model with vertical and horizontal wind components, microphysical quantities, perturbation potential temperature, geopotential and surface pressure of dry air as prognostic variables. The model was configured with three two way nested domains with grid spacing of 27, 9 and 3 km centered over central Nepal (26.34° N, 83.12° E), which was defined in the Lambert conformal projection as shown in Figure 4.1. The coarse domain (d01) covered almost the entire Hindu – Kush Himalaya region, the Indian subcontinent and the Bay of Bengal to simulate regional flow patterns. The domain 2 (d02) covered all of the Nepal and the innermost high resolution domain (d03) covers central Nepal including the Kathmandu valley. The innermost domain (3 km x 3 km), which contains 111 x 111 horizontal grid points and 40 vertical levels with the domain top pressure of 50 mb, is used to evaluate the model results. The model integration time-step was set to 30 s. We used NCEP/DOE reanalysis 2 data for model initialisation and boundary conditions. Twenty four categories of terrestrial data were obtained from the USGS ‘30s’ global data set ([http://www.mmm.ucar.edu/wrf/users/download/get_source2.html](http://www.mmm.ucar.edu/wrf/users/download/get_source2.html)). The model was initiated at 09 UTC on 17 May 2011 and ran for 39 hours.

The physics packages used in the simulations comprise the Dudhia short-wave radiation (Dudhia, 1989), the RRTM long-wave radiation (Mlawer et al., 1997), the YSU boundary layer and the Noah land surface models (Ek et al., 2003). As the model explicitly resolves the cumulus convection in the high resolution simulation (3km x 3km) hence no cumulus parameterization was used in the innermost domain however, it was used in the other coarse resolution domains. More details of the model can be found in the WRF technical report by Skamarock et al. (2008) and the WRF Users Guide (Wang et al., 2012). Brief description of microphysical parameterizations to be applied here in this study is given below.
Figure 4.1: Domains used for WRF simulation with horizontal grid resolution 27 km (domain 1), 9 km (domain 2) and 3 km (domain 3) centered over central Nepal (26.34° N, 83.12° E) and the model is set to 40 vertical levels.

The model experiment designed here in this study is a performance evaluation of the four microphysical schemes, generally with an increasing complexity of parameterizations, to simulate a convective storm that generally occurs during the pre-monsoon season (March – May) in Nepal. These microphysics schemes were tested to examine how precipitation evolution depends on treatment of moisture processes in the parameterization. The schemes are as follows: WRF single moment 6-class scheme – WSM6 (Hong and Lim, 2006; Hong et al., 2004), WRF double moment 6-class scheme – WDM6 (Lim and Hong, 2010), Lin scheme (Lin et al., 1983; Rutledge and Hobbs, 1984) and Morrison double moment scheme (Morrison and Pinto, 2005; Morrison et al., 2005; Morrison et al., 2009). The simulated results from the four microphysical schemes were then compared with the limited in-situ observations obtained from our field experiment at Nagarkot, Nepal. Details of the observations are described in the subsequent sections.
1) Lin scheme: This microphysical parameterization scheme includes 6-class water species (e.g. water vapor, cloud water, rain water, cloud ice, snow and graupel) as prognostic variables. This is a single moment bulk microphysics scheme which predicts mixing ratio of the water species and assumes exponential size distribution of the hydrometeor particles. Cloud water and ice particles are assumed to be mono-disperse and small enough to neglect terminal velocity. Ice crystal formation is considered to be temperature dependent. More details of the scheme can also be found in Chen and Sun (2002).

2) WSM6 scheme: This is a mixed-phase cloud microphysics parameterization scheme, which predict mixing ratio of the 6-class water substance variables. The warm phase cloud processes in this scheme are quite similar to the aforementioned Lin scheme however it differs in the treatment of ice cloud parameterizations. In addition to the large hydrometeors particles, the scheme also includes gravitational sedimentation of cloud ice.

3) WDM6 scheme: This scheme is quite similar to the WSM6 scheme as it consist the same prognostic water substance variables. However, this is a double moment scheme for the warm-phase cloud hence it predicts both mixing ratio and number concentration of the warm-phase hydrometeor variables (i.e. cloud water and rain water) including number concentration of cloud condensation nuclei (CCN).

4) Morrison double moment scheme: This is a double moment mixed-phase cloud microphysics parameterization scheme which predicts mixing ratio and number concentration of the 6-class water substance variables. Cloud droplet size distribution is explained by the gamma distribution and the other remaining hydrometeor particles are assumed to follow the Marshall – Palmer size distribution (i.e. inverse exponential distribution). As the scheme does not explicitly predict CCN concentration, aerosols concentration and characteristics can be prescribed. Droplet activation occurs as a function of grid and sub-grid scale vertical velocity.
3. Results and discussion

3.1 Details of the convective storm

The convective storm occurred in the late afternoon of 18 May, 2011. The infrared and visible spectrum of geostationary satellite images are depicted in Figure 4.2(a-d), which shows synoptic weather conditions associated with the convective storm. The upper two figures (a-b) are received at 00 UTC 18 May 2011 approximately 9 hours before the storm started. These images showed that dense convective clouds formed over east Nepal and northeast India covering a wide geographical area. Over the next 12 hours (i.e. 1200 UTC 18 May 2011, during the storm) the satellite images as shown in the lower two figures (c-d), revealed that the clouds moved over central Nepal with a significant reduction in its size. It is sensible that the clouds may have produced heavy rainfall across east Nepal and northeast India subsequently loses its strength and size. The rainfall in central Nepal was observed by a Vaisala weather sensor installed at the Nagarkot, Nepal (Lat: 27.7°N, Lon: 85.5°E, Alt: 1900m). The observed features of surface meteorology are described in section 3.2 which shows dry bulb temperature, relative humidity, wind speed, wind direction, rainfall, and surface pressure.
Figure 4.2: Geostationary satellite (MET7) images (a) infrared, and (b) visible spectrum received at 00 UTC 18 May 2011 (5:45 Local) (c) infrared, and (d) visible spectrum received at 1200 UTC 18 May 2011 (17:45 Local).

3.2 Simulation of surface meteorology

Figure 4.3(a-b) shows a comparison of observed and the WRF simulated surface temperature and relative humidity (RH) at Nagarkot, Nepal. Our simulations show that the temperature is not very sensitive to the microphysics parameterization and it is generally overestimated. However, the simulations were able to reproduce diurnal cycle of surface temperature. Both observed and simulated temperature patterns show a drastic drop in the environmental temperature as the rainfall started, which is attributed to the evaporative cooling mechanism because when the raindrops fall below the saturated layer, air cools from evaporation of rain and melting of hail particles. The cooling is also possible due to passage of cold fronts ahead of the storm. Rapid cooling of environment with a magnitude of more than 7°C per hour was found due to passage of cold front ahead of the thunderstorm in Southeast India (Rajeevan et al., 2010). Like the surface temperature, simulations show that the RH is also not very sensitive to the microphysics parameterizations, and it is normally underestimated.
Figure 4.3: Time series analysis of observed and simulated for the four microphysics schemes (a) surface temperature, and (b) relative humidity at Nagarkot, Nepal.

Figure 4.4(a-b) shows that there is a weak sensitivity of wind speed and wind direction to the microphysics parameterizations. Observed patterns of these variables measured by the Vaisala meteorological sensor are in good agreement with the simulated patterns. An abruptly increase in wind speed and sudden change in wind direction from easterly to south-westerly was observed before the storm onset that transported moisture laden air from the Bay of Bengal to the measurement site at Nagarkot.
Figure 4.4: Same as Figure 4.3 but for (a) wind speed, (b) wind direction, and (c) Histogram of rainfall comparing different microphysical schemes.
Figure 4.5(a) shows that the simulated surface pressure is not sensitive to the microphysical parameterizations and it is greatly overestimated as compared to the observation. We noticed that in both observed and simulated cases the surface pressure started to drop long time before the storm initiated and remained low until the storm dissipated, which may be associated with development of clouds. Dai and Trenberth (2004) found that surface pressure is overestimated by 20–50% over low-latitude and underestimated by the same amount over mid-latitude in their model performance evaluation.

A strong sensitivity of rainfall to cloud microphysical parameterizations was observed in our simulations Figure 4.5(b), although it underestimated the actual rainfall intensity and completely missed out some earlier episodes of the rainfall. We found that the Morrison scheme produced 72% less rain than the observation. Similarly, the Lin, WSM6 and WDM6 schemes generated 93%, 86% and 95% less rain than the observation respectively also shown by histogram (Figure 4.4(c)). It seems that none of the parameterization schemes are able to reproduce the observed patterns of rainfall. However, among the four different microphysical schemes used in the simulations, the Morrison double moment scheme performed best. It did not show any statistically significant difference compared to the observed rainfall at the 80% confidence level using a Chi-squared goodness of best fit.

This forecast error may be attributed to the grid resolution (3km x 3km) considered here in this study, which may not be able to simulate very detail features of the storm. Petch (2006) carried out a sensitivity study of convection development to model grid resolution using the Met office cloud-resolving model. The author highlighted that initiation of convection is significantly delayed in coarse grid resolution (>200m) and rapid growth of cloud was observed immediately after initiation of convection, however such effects were not noticed when horizontal grid resolution was reduced to less than 200m.

The error may also be attributed to the coarse lateral boundary conditions used in this study, which was interpolated from 2.5° x 2.5° horizontal resolution and 17 pressure levels from the NCEP reanalysis data. The MM5 simulated
rainfall was considerably underestimated across the continental United States as revealed by a cloud microphysical sensitivity analysis of the warm season convection (Liu and Moncrieff, 2007). They argued that the forecast error was caused by the coarse grid resolution (3km) used in their study and the lateral boundary condition (40km) obtained from the NCEP Eta model analysis.

**Figure 4.5**: Same as Figure 4.3 but for (a) surface pressure, and (b) rainfall.

A spatial distribution of 39 hours accumulated rainfall for the four microphysical parameterization schemes are shown in Figure 4.6(a-d). All the simulations generally produced WNW – ESE oriented precipitation corridor propagating toward northwest direction over time (not shown). Genesis of the propagating convection was the eastern foothills of the Himalaya, which subsequently moved toward central Nepal, lost its original strength and size as
shown in the geostationary satellite images, attributed to deposition of heavily moist and warm water vapor as precipitation across the eastern foothills of the Himalayas.
Our simulations showed a quite similar pattern of heavy rainfall that occurred across south-eastern part of the domain as estimated by the BMPs schemes. Furthermore, it was observed that the windward slope of the hills received more rain, which is confined across the ‘Terai’ (low land in South of Nepal, altitude < 200m), Siwalik and Mahabharat ranges (geographical regions
having 10 – 50 km wide swath with altitude ranging from 200 – 3000 m extending parallel to south of the Himalayas), suggested a dominant role of topography. However, there are few discrepancies among the cloud microphysical parameterization schemes. For example, the Lin scheme produced a number of organized rain bands. In contrast, the Morrison scheme showed a uniform spatial distribution of rainfall as compared to the other microphysical schemes. On average, the Morrison scheme produced 584 mm rain across the domain during 39 hours of simulation. Similarly, the Lin, WSM6 and WDM6 scheme generated 667 mm, 585 mm and 495 mm rain respectively.

3.3 Simulation of vertical velocity

Figure 4.7(a-d) shows evolution of vertical velocity over time and height at Nagarkot simulated for the four microphysics schemes. As convective storm is characterized by a strong updraft and downdraft, here we want to show whether or not the simulated vertical velocity is able to explain the observed rainfall patterns. A back-of-the-envelope comparison of the rainfall occurrence time and the vertical wind speed at Nagarkot explained that the model was able to capture the convective motion. All the four microphysical parameterization schemes produced strong convective motion during beginning and near end of the simulations, which was fully explained by the observed rainfall at Nagarkot. Although the schemes captured the updraft, surprisingly, none of the schemes produced rainfall in beginning of the simulation; again this may be attributed to the coarse grid resolution (Petch, 2006). The Morrison scheme produced stronger updraft (more than 2m/s) and downdraft (~1.5m/s) than the other schemes, which subsequently generated more rain than its counterparts. We observed that the updraft cores could reach up to 15 km high as shown in the simulations.
Figure 4.7: Vertical velocity (m s$^{-1}$) at Nagarkot, simulated for the four microphysical parameterization schemes (a) Lin, (b) Morrison, (c) WDM6, and (d) WSM6 scheme.

Figure 4.8(a-d) shows time – latitude diagrams of the simulated vertical velocity for the four microphysics schemes at a mean height of ~5km from the ground level, which reveals a moderate sensitivity to the chosen parameterization scheme. All the schemes showed a northward propagation of convection with strong updraft and downdraft in the beginning and near end of the simulations. The updraft track showed a strong dependency of convection on topography because strong updrafts were observed when the storm moved towards the high elevation region (i.e. northward).
Figure 4.8: Time – Latitude diagrams for the vertical velocity at a mean height of ~5km from the ground level simulated for (a) Lin, (b) Morrison, (c) WDM6, and (d) WSM6 scheme.
3.4 Simulation of hydrometeor profile

Figure 4.9: Domain and time averaged (a) total water condensate (g kg\(^{-1}\)), and (b) cloud fraction (%) simulated for the four microphysical parameterization schemes.

Figure 4.9(a-b) depicts domain and time averaged vertical profile of total water condensate and cloud fraction simulated for the four microphysical schemes. The total water condensate of a grid cell was calculated by adding up mixing ratio of the hydrometeors (cloud water, cloud ice, rain water, snow and graupel) in the grid cell. The cloud fraction calculation is adopted from Liu and Moncrieff (2007), based on mixing ratio of hydrometeors which assumes 100% cloudiness over a grid box when sum of cloud water, ice and snow mixing ratio exceeds 0.01g/kg.

A quite similar total water condensate distribution among the parameterization schemes was found in the lower troposphere (i.e. below 600hpa) however, significant difference was observed in the upper troposphere. The minimum condensate was produced in the Lin scheme throughout the atmosphere. In contrast, the WDM6 scheme produced highest total condensate in the middle of the atmosphere. These effects were attributed to the different
assumptions made in the cloud microphysical parameterization processes, for example, the schemes use different density of graupel and also differ in the intercept parameters estimation of hydrometeors and ice phase processes. Moreover, the sensitivity may also have been caused by moment of distribution of the hydrometeors (e.g. single or double moment).

In the cloud fraction, again the Lin scheme showed minimum clouds in the atmosphere, in contrast, the Morrison scheme produced more high level clouds (above 500hpa) and less low level clouds. The WSM6 and WDM6 schemes produce more low level clouds. This cloud cover variability is attributed to the different approach of parameterizations used in the schemes such as ice sedimentation, collection efficiency etc… For example, a high collection efficiency of ice, which in fact depends on temperature, cause fast transformation of cloud ice to snow then subsequently precipitate out due to its large sedimentation velocity. Such effect was observed in the study of deep tropical convection, where the WRF simulated anvil clouds for the Morrison microphysics parameterization scheme was found less persistent than in the observation (Connolly et al., 2013).
Figure 4.10: Domain and time averaged vertical profile of cloud water (thick dashed), rainwater (dashed), ice (thick dotted), snow (dotted), graupel (solid) and total condensate (thick solid) mixing ratio (g kg\(^{-1}\)) simulated for (a) Lin, (b) Morrison, (c) WDM6, and (d) WSM6 scheme.

Figure 4.10(a-d) shows domain and time averaged vertical profiles of individual hydrometeor simulated for the four microphysics parameterization schemes. The simulations showed a noticeable variation in the vertical profiles and more pronounced effects were observed in the ice phase. The warm-phase hydrometeors (i.e. cloud water and rain water) distribution exhibited several common features. In contrast, the snow mixing ratio in the Morrison scheme was several orders of magnitude more than in the other schemes. The WSM6 produced slightly more snow in the upper troposphere than its counterpart WDM6.

The Morrison scheme showed less graupel mixing ratio than in the other three schemes, which may have been offset by the snow mixing ratio. A quite similar graupel mixing ratio profile was observed in the remaining three schemes (i.e. Lin, WSM6 and WDM6). A number of other studies have also highlighted
the significant variation of hydrometeors vertical profile in cloud microphysical sensitivity analysis (Liu and Moncrieff, 2007; Liu et al., 2011; Rajeevan et al., 2010).

4. Summary

In this study we carried out a microphysical sensitivity of the high resolution (3km x 3km) WRF model to simulate a convective storm that developed over central Nepal. The study domain is characterized by the complex Himalayan terrain in north including the Mt. Everest (8,848 m) and ‘Terai’ (low land in South of Nepal, altitude < 200m) in south. The overall objective of this paper is to analyze performance evaluation of different microphysics schemes available within the WRF model.

We used four cloud microphysical schemes (Lin, WSM6, WDM6 and Morrison), generally parameterized with an increasing complexity, from single to double moments of particle distribution. Furthermore, the model results were also compared with the limited observation that obtained from the field experiment at Nagarkot, Nepal (Lat: 27.7°N, Lon: 85.5° E, Alt: 1900m). The convective event occurred in May 18, 2011 which is just before onset of the summer monsoon in Nepal, characterized by an unstable atmosphere with a hot and dry weather condition. Note that the conclusions drawn here in this study are entirely based on analysis of single convective event, which may not be a representative storm of the region. The evaluation is based on the time series analysis of surface meteorology, vertical profiles of hydrometeors, updraft speed and cloud cover. The major results are summarized as below:

- Surface rainfall is underestimated by the bulk microphysical parameterization schemes used in this study. The Morrison scheme generated 72% less rainfall than the observed amount. Similarly the Lin, WSM6 and WDM6 produced 93%, 86% and 95% less rainfall respectively. The discrepancies between simulated and observed rainfall may be attributed to the parameterizations of cold cloud microphysics,
particularly snow, which has highly varied habit types, densities, and fall velocities than the other precipitation species (e.g. rain and graupel) that significantly affects microphysical process and these diverse properties of snow are not well represented in the model. However, the Morrison scheme performed better than the other schemes which did not show any statistically significant difference compared to the observed rainfall at the 80% confidence level using a Chi-squared goodness of best fit. The domain averaged 39 hours accumulated rainfall is 584mm, 667mm, 585mm and 495mm for the Morrison, Lin, WSM6 and WDM6 schemes respectively.

- The convective storm, transports moist air from the Bay of Bengal, generally moves from southeast to northwest direction, encounter the giant Himalayan barrier which in turn dump most of its water vapor as precipitation and loses its strength. The heavy precipitation was found to be confined across the ‘Terai’, Siwalik and Mahabharat ranges (geographical regions having 10 – 50 km wide swath with altitude ranging from 200 – 3000m extending parallel to south of the Himalayas). Our simulations show that the windward slope of the hills receives heavy precipitation. Such mechanisms are not uncommon in the mountainous terrain (Barros and Lettenmaier, 1994) because moist air is forced up to high elevation which in turn condensed and produce precipitation along the windward slope of the topographic barrier.

- A strong sensitivity of ice phase hydrometeors and cloud cover to microphysics were observed. The ice mixing ratio in the WSM6 and WDM6 were greater than in the Lin and Morrison schemes. In contrast, the Morrison scheme produced highest snow mixing ratio, whereas graupel content was the lowest. The scheme also generated maximum cloudiness in the upper and minimum in the lower troposphere.

- The surface pressure showed a weak sensitivity to the microphysical parameterizations used in the model and it was significantly
overestimated, although it followed the diurnal cycle as reported in the observation. The relative humidity showed a moderate sensitivity to microphysical parameterizations.

- The vertical wind speed showed a moderate sensitivity to microphysics parameterization. The double moment microphysics scheme showed a stronger updraft speed than the single moment schemes. All the parameterization schemes captured propagation of the storm reasonably well, and were also able to simulate the topographic effects because an increased updraft speed was observed when the storm moved northward.

To make a general statement from the aforementioned conclusions would be sceptical as it is derived from single convective event. It may be more effective to carry out simulations of several other convective events, covering different synoptic conditions and considering other physical processes such as parameterized cumulus convection, boundary layer and land surface schemes. A strong sensitivity of the model to the parameterized cumulus convection than the explicit convection was found across the continental United States (Liu et al., 2006). Convection permitting simulation configured with a single domain could produce better results than multiple domains with parameterized convection in the coarse and explicit convection in the high resolution domain because the parameterized convection in the outer domain could influence inner domains (Liu and Moncrieff, 2007; Rajeevan et al., 2010), however the single domain approach could enhance errors while feeding coarse resolution data from a global model directly to the high resolution grid (Leung and Qian, 2003). Cloud microphysics strongly interrelated to dynamics, on the other hand the dynamics is influenced by topography and associated gravity waves, which can not be neglected where topography plays vital role such as in the Himalayas. Hence, parameterization of these effects could improve our understanding of cloud microphysics in the Mountainous terrain.

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References


CHAPTER 5: Sensitivity of Indian monsoon precipitation to aerosols and temperature perturbations over the foothills of the Nepal Himalayas

The following chapter has been submitted (under review) to Atmospheric Research for publication.

Authors: Rudra K. Shrestha¹,², Paul J. Connolly¹, and Martin W. Gallagher¹

¹School of Earth, Atmospheric and Environmental Sciences, The University of Manchester, UK.
²Sustainable Consumption Institute (SCI), The University of Manchester, UK

Contribution from others: Paul Connolly helped with the initial model set up and data analysis and also reviewed and commented on the paper. Martin Gallagher reviewed the paper and gave critical comments. All work was carried out under the supervision of Martin Gallagher and Paul Connolly.

Abstract

Increasing anthropogenic aerosol emissions in the Nepal region may lead to a modification of regional cloud properties, there by altering the balance between cloud water phase and cloud height, which may lead to an alteration of spatial precipitation patterns within the Kathmandu valley.

Climate change induced temperature enhancement across this region may also influence cloud macro- and micro-physical properties by altering the available water budget. In this paper, we simulate the possible effects of aerosol and temperature perturbations on the evolution of precipitation and its spatial distribution during the Indian Monsoon season over the Himalayas in this region.
We present results using the Weather Research and Forecasting (WRF) model, coupled with a bulk microphysics scheme, in a convection permitting configuration. The sensitivity of rainfall to changes in background aerosol and temperature is analyzed for three real-world case studies, which occurred during different seasons of the year, having varying rainfall intensity. Furthermore, in order to assess approximately how aerosols may affect regional Nepalese precipitation in the future, we compare results from the control run with those where the temperature profiles are perturbed both i) uniformly, and ii) randomly for three aerosol loading cases; corresponding to ‘low’, ‘medium’ and ‘high’ concentration based on current observations and realistic extrapolated values. In this study we also use a more realistic prognostic CCN approach to investigate the effects of aerosol on regional precipitation.

In summary we found that rainfall is only marginally sensitive to the range of aerosol loadings investigated. The effect, however, is nonlinear, ranging from -3% to +4% depending on the conditions of the simulated case. The model results highlight a 1st indirect (Twomey) effect reasonably well; however, somewhat surprisingly, rainfall was not particularly sensitive to the aerosol perturbations used. The model showed that the distribution of high ice mass present in the upper troposphere and the associated ice phase processes appear to play a crucial role in buffering the sensitivity to increased aerosol loading. We do observe that aerosol perturbations can modify the shape, size and spatial distribution of individual cloud regions and their precipitation production. However, for the runs performed here, the effect of temperature perturbations is more than the aerosol effect, ranging from -17% to +22% and +31% to +93% in the random and uniform temperature perturbation cases respectively, which suggest more intense rain events are likely as the climate warms in this region.
1. Introduction

People living in South Asia are dependent upon monsoon precipitation for their daily lives (Mooley and Parthasarathy, 1984; Turner and Annamalai, 2012) since more than 70% of the working population rely on agriculture (Kumar et al., 2004). The majority of agricultural systems in this region are rainfall dependent. Accurate prediction of the monsoon onset and decay, its movement and its variability across the region is an important problem which subsequently influences the national agenda of the region in the areas of sustainable development, poverty reduction and disaster management.

An assessment of the Intergovernmental Panel on Climate Change (IPCC) concluded that the global mean temperature in the fossil-intensive (A1FI) scenario, often thought of as a ‘worst case’, will produce a rise in the range 2.4°C to 6.4°C by the end of 21st century (IPCC, 2007a). IPCC (2007b) emphasizes that the A1B scenario, which is considered as a moderate emission scenario, generally projects lower temperature than the A1FI, however it predicts nearly the same magnitude as in the A1FI scenario (i.e. up to 6.1°C) across the Tibetan Plateau and a further rise is expected during the dry season (i.e. up to 6.9°C). A number of studies have highlighted that sensitivity of the Himalayan region to these perturbations and show that it is warming faster than other regions of the globe (Liu and Chen, 2000; Shrestha et al., 1999). A general rise in temperature increases the water holding capacity of the air, (the ‘Clausius – Clapeyron effect’), which in turn may increase the potential for heavy precipitation in convective or orographic initiated storms.

The aerosol loading in this region is hypothesized to significantly affect aerosol-cloud and precipitation interactions as increasing aerosols will act as a source of additional cloud condensation nuclei (CCN). In warm stratus clouds, increased aerosols may subsequently decrease the average cloud droplet effective radius resulting in a negative radiative forcing – the first indirect effect (Twomey, 1977). This consequently suppresses rainfall for a given water path – the second indirect effect for warm clouds (Albrecht, 1989).
Recently Li et al. (2011) and Connolly et al. (2013) have pointed out that increasing aerosol loading may either enhance or suppress cloud dynamical development and the production of precipitation. These potential perturbations are of major concern across the south Asia region as air pollution in the region has significantly increased over the past years due to population increases and rapid economic growth and this is set to continue. Further studies aiming to understand how aerosols, clouds and precipitation interact in this region are therefore warranted, and this remains the major challenge to the climate-weather modelling community (NSF, 2011). Understanding the monsoon interaction in this region is complicated (Turner and Annamalai, 2012) as it is controlled by the rugged and complex topography of the Himalayas (Wang, 2006), its interaction with large scale dynamics, and cloud microphysics. This causes wide ranging effects on pollution transport, clouds and precipitation formation and their distribution (Houze et al., 2007).

Shrestha et al. (2010) suggest that advected aerosol pollution from the Indo-Gangetic Plains is confined in the many deep valleys, localizing their spatial distribution along the foothills of the Himalayas, and this can play a significant role in modulating cloud microphysical processes and the distribution of orographic precipitation. This same effect was observed in the Sierra Nevada Mountains by Lynn et al. (2007) where ‘low’ concentrations of aerosols produced more precipitation in the foothills of the Mountains and ‘high’ aerosol concentrations produced less precipitation, but resulted in maximum precipitation levels shifting to the leeward slopes in the ‘low’ aerosol concentration regime. Muhlbauer and Lohmann (2008) pointed out that for orographic clouds over the Swiss Alps (e.g. in the region of the Jungfraujoch mountain), increasing aerosol number concentrations could suppress the contribution of warm phase cloud processes, leading to a decrease in orographic precipitation on the upslope side of the mountains there. In the mixed-phase cloud regime orographic precipitation depends on a knowledge of the detailed properties of the ice phase that evolves within the cloud (Muhlbauer and Lohmann, 2009). However, in a similar study, aerosol effects on precipitation generally ranged from -19% to 0% in the case of stratiform orographic mixed-phase clouds (Muhlbauer et al., 2010). Connolly et
al. (2013) performed a detailed sensitivity study of aerosols in deep tropical convective clouds using the ‘high’, ‘medium’, and ‘low’ background aerosol concentration approach and found a weak effect on precipitation. However these effects were very nonlinear and no general relations between aerosols and precipitation were found because of interactions between colliding gust fronts. A weak sensitivity of precipitation to aerosol perturbation was observed by Lee and Feingold (2010) in the simulation of a tropical convective cloud system. They found that 10-fold increase in aerosol concentration increases nearly 9% precipitation, which is caused by the compensating effects of enhanced convective rain to suppressed stratiform rain.

In addition to their contribution to CCN effects, anthropogenic aerosols are identified as one of the major sources of Ice Nuclei (IN), and so can influence several other ‘indirect effects’ as suggested by Lohmann and Feichter (2005). These include the ‘thermodynamic’, ‘glaciation’ and ‘riming’ indirect effects which may modulate mixed-phase properties of clouds; however, the impact of these effects on precipitation are still quite uncertain although in general it is thought that increasing IN levels in mid-level clouds will lead to increased ice mass leading to a general reduction in net cooling from such clouds on climate. Lohmann and Feichter suggest that increasing aerosol number concentration will decrease the riming efficiency occurring in these clouds; however, Connolly et al. (2006) argued that enhanced aerosol concentrations in deeper convective clouds may actually accelerate the ‘riming’ process as there is an association with more liquid water being present at higher altitudes. This tended to produce more precipitation during the dissipation stage of a storm, so that although the timing of precipitation was affected there was little change in the precipitation amount.

A number of field experiments have reported aerosol mass concentration across the Himalayas which show considerable seasonal variations, with a maximum occurring during the pre-monsoon and a minimum during the monsoon season (Bonasoni et al., 2008; Marioni et al., 2010). This suggests a potentially strong interaction between the summer monsoon and the mass concentration of aerosol particles in particular. Although aerosol number concentrations are an important parameter to investigate aerosol - cloud - precipitation interactions due
to their ability to act as CCN or IN, very limited field experiments have been
carried out to measure the number distribution of aerosol in the Himalaya region
and none have investigated cloud properties by in situ methods there. Shrestha et
al. (2010) reported aerosol number concentrations during the pre-monsoon season
from two locations in the middle mountain region of Nepal, which may typically
represent urban and rural sites. They found average total particle concentrations in
the size range 30 – 340 nm were 5, 334 ± 3,355 per cm$^3$ and 7,566 ± 8,013 per
cm$^3$ for “urban” and “rural” sites respectively with strong diurnal cycles peaking
in the morning and evening, although the higher concentration at the latter was
affected by strong local sources linked to transport induced by local topography.

Modelling experiments by Meehl et al. (2008) reported significant
increased heating of the lower troposphere across South Asia due to absorbing
aerosols leading to the so-called ‘elevated heat pump’ theory. Warming of the
Tibetan Plateau during the pre-monsoon season (March – May) has a positive
effect on the Indian summer monsoon precipitation in the study of Meehl et al..
However, the model studies of Bollasina et al. (2011) suggest the opposite: that
there is a decreasing trend of Indian summer monsoon precipitation. Their
simulations showed aerosols suppress the land – ocean temperature contrast,
which in-turn slows down the inter-hemispheric meridional overturning
circulation causing a weakening of the Indian summer monsoon precipitation.
Nevertheless, aerosols have been shown to have both a warming and cooling
effect and a complete understanding of resultant impact of greenhouse warming
and aerosols effects in this region is still incomplete.

In this paper we aim to investigate how aerosol and temperature
perturbations may affect cloud microphysical processes and to what extent
subsequent aerosol - cloud interactions may alter precipitation formation and
distribution over the complex terrain of the Himalayas. We address the following
questions:

• How will the precipitation respond to aerosol and temperature
  perturbations?
• How is the spatial distribution of the precipitation affected by aerosol perturbations?
• How are these perturbations manifested in the microphysical properties of the clouds?

This paper is organized as follows: the regional climatology of the model domain will be described in section 2; the numerical model and microphysics are explained in section 3 and 4 respectively; Section 5 outlines the experimental set up and the results follow in section 6. A discussion and conclusion is presented in section 7 and finally limitation of this study is discussed in section 8.

2. General climatology of the region

The study domain is characterized by complex terrain, with altitude varying from 70 m in South eastern Nepal, to the highest peak in the world, 8850 m (Mt. Everest) in the North, forming a sharp North - South topographical gradient and which extends from the West to the East of the region covering northern India, Nepal and the Tibetan Plateau.

The Indian summer monsoon, which is also called the southwest monsoon, is influenced by the large Himalayan barrier which forces the monsoon air current to ascend, subsequently producing clouds and precipitation. It is different from the other monsoon systems around the globe due to the prominent role of the regions massive topographical features (Cadet, 1979). The Indian summer monsoon is generally most intense during July through September and gradually weakens over time, subsequently dissipating by the first week of October. Barros and Lang (2003), Bollasina et al. (2002) and Bonasoni et al. (2008) show that summer monsoon onset across the foothills of the Nepalese Himalayas (e.g. Khumbu valley and Marsyandi basin) generally occurs in mid-June and fades out by the beginning of October. The monsoon rainfall is attributed to moist south-westerly flow originating from the Bay of Bengal during summer time. The monsoon season (June - October) is responsible for 80% of the annual rainfall (Shrestha, 2000) in the region, and is preceded in the pre-monsoon season (March - May) by
violent lightning and thunderstorms (Barros and Lang, 2003). Significant variability in the summer monsoon precipitation is observed across the foothills of the Himalayas with more rain in the south-east of Nepal and less precipitation across the north-west of the country. This is generally attributed to convergence of moist air while moving to the region. As a result total summer monsoon rainfall can exceed 350 - 450 cm in central Nepal (Barros et al., 2000; Lang and Barros, 2002; Shrestha, 2000).

The winter monsoon (December - February), also known as the northeast monsoon, is attributed to the north-easterly flow which carries little moisture, causes short episodes of rainfall over low land but causes heavy snow fall over the high altitude region. Usually a western disturbance, such as an extra-tropical cyclone, originating in the Mediterranean, brings sudden non-monsoonal winter precipitation when the storm system encounters the Himalayas. In contrast to the summer monsoon flow, the western disturbance moves from west to east over the region. However, it is not explicitly understood how the western disturbance intensifies the winter precipitation across the Nepalese Himalayas (Lang and Barros, 2004).

3. Numerical model

The simulations described here were carried out using version 3.1.1 of the Weather Research and Forecasting (WRF) model. The model incorporates the advanced Research WRF (ARW) mass based, terrain following vertical coordinate system, with adjustable vertical grid spacing. Prognostic variables include vertical and horizontal wind components, microphysical quantities, perturbation potential temperature, geopotential and surface pressure of dry air. Twenty four categories from the USGS ‘30s’ global data set were used to initialize surface properties such as terrain, vegetation index and land use type.

The model uses a 3rd order Runge - Kutta time integration scheme with 5th order horizontal and 3rd order vertical momentum advection options. A gravity wave absorbing layer was used to damp anomalously large vertical velocities in
the model (Klemp et al., 2008). A complete description of the WRF model is provided by Skamarock et al. (2008).

The WRF model has been widely tested and successfully applied for the evolution of mesoscale weather phenomenon in areas of complex topography (Jimenez et al., 2010). Maussion et al. (2010) for example applied the WRF model to simulate a precipitation event over the Tibetan Plateau and compared predictions with the Tropical Rainfall Measuring Mission (TRMM) satellite observations. The simulated precipitation was found to be in good agreement with the satellite observation, giving confidence that our set-up will also be suitable for this study.

4. Cloud Microphysical Description

The Morrison double-moment bulk cloud microphysical scheme (Morrison et al., 2005a; Morrison et al., 2009) was implemented in WRF in this study. This double moment microphysics scheme thus predicts the number concentration and the mixing ratio of different hydrometeors. Prognostic equations are used to describe the evolution of five hydrometeor species (cloud droplets, rain, cloud ice, snow and graupel) and water vapour. CCN activation is parameterized following the standard Twomey (1959) power-law relation, $N_i = C_{S_i}^{k}$, using the modification described by Rogers and Yau (1989) for use in this model. This scheme does not explicitly include calculation of peak supersaturation, since this is not resolved by the model grid, but instead it calculates the number of activated droplets as a function of updraft velocity. The maximum super-saturation is therefore implicit in this formulation, as it mainly depends on updraft speed and the CCN characteristics ($C$ and $k$). As in other approaches using this power-law formulation, there is no maximum limit to the nuclei that are activated. The scheme does not explicitly use aerosol or CCN concentration, and assumes an a-priori background value. Following Twomey (1959), the equation governing the number of CCN activated at cloud base is:

$$N = 0.88C^{2/(k+2)}\left[7\times10^{-2} \times U^{3/2}\right]^{1/(k+2)}$$
where $N$ describes the total number of activated droplets ($\text{cm}^{-3}$). The vertical velocity, which includes grid and sub-grid scale velocity, is represented by $U$ (cm s$^{-1}$), $C$ is the background CCN number concentration ($\text{cm}^{-3}$), and $k$ represents an activation parameter that depends on air mass type and which may vary from 0.4 to 1.0. In this study we use a constant value of $k$ equal to 0.8.

In WRF the warm rain process is parameterized using the method of Khairoutdinov and Kogan (2000) which was arrived at by regression analysis against droplet size spectra predicted by an explicit bin-microphysical model. The ice phase descriptions in the model include diffusional growth, aggregation, riming and melting of ice hydrometeors (processes relevant for cloud ice crystals, snow and graupel). Heterogeneous ice nucleation was parameterized following Rasmussen et al. (2002), where primary ice crystals are generated through immersion freezing, contact freezing and deposition and condensation nucleation. Ice multiplication processes were parameterized following Hallett and Mossop (1974).

5. Model setup

The model was configured using three (two-way) nested domains with a horizontal grid resolution of 27, 9 and 3 km centred over central Nepal ($26.34^\circ$ N, $83.12^\circ$ E) defined in the Lambert conformal projection (Figure 5.1). The outermost domain (Domain 1) covered the Asian monsoon region including the Himalaya range, the Indian subcontinent and the Bay of Bengal to capture regional flow patterns. Domain 2 covered all of the Nepal region including the central Himalayas and part of Northern India in order to capture the key mesoscale circulation patterns there. The innermost high resolution domain (Domain 3) covers central Nepal focusing on circulation over the complex terrain of the Himalayas itself. Evaluation of the model results is focused on the high resolution (3km x 3km) domain which contains 111 x 111 horizontal grid points and 40 vertical levels with the domain top pressure of 50 mb. The model integration time-step was set to 30 s. We use NCEP/DOE reanalysis 2 data for model initialisation and boundary conditions. The physics packages include the
Dudhia short-wave radiation (Dudhia, 1989), the RRTM long-wave radiation (Mlawer et al., 1997), the YSU boundary layer (Hong et al., 2006) and the Noah land surface models. No cumulus parameterization was used in the high resolution (3km x 3km) simulations presented here as the model set up should explicitly resolve the cumulus convection in sufficient detail, but it was used in the lower resolution domains.

The atmospheric moisture content was altered by perturbing the temperature i) uniformly and ii) randomly across the domain by modifying the model standard code and holding the relative humidity constant. In the uniform perturbation case a constant increment is added to the unperturbed temperature (control, hereafter Tctrl) at every model grid point. Two different temperature enhancement scenarios i) Tctrl+5, and ii) Tctrl+10 were created with respect to the Tctrl run by adding $\Delta T = 5^\circ C$ and $10^\circ C$ respectively. In contrast, the random perturbation was created using a random number generator. The range of perturbations was restricted to between $\Delta T = -0.5^\circ C$ to $+0.5^\circ C$, $\Delta T = -2.5^\circ C$ to $+2.5^\circ C$ and $\Delta T = -5.0^\circ C$ to $+5.0^\circ C$, forming three different temperature perturbation scenarios relative to the control simulation, i.e., i) Tctrl±0.5, ii) Tctrl±2.5, and iii) Tctrl±5 in order to investigate sensitivity to temperature perturbation in the forecast at different background aerosol concentrations. The aerosol concentrations were prescribed as 500 cm$^{-3}$, 1,500 cm$^{-3}$ and 3,500 cm$^{-3}$ referred to as the ‘low’, ‘medium’ and ‘high’ aerosol cases respectively. These aerosol perturbation scenarios provide only approximate ranges for the region (Shrestha et al., 2010), with the high aerosol case being considered as biomass burning event for example. However these need to be verified with further observations.
Figure 5.1: Domains used for WRF simulation with horizontal grid resolution 27 km (domain 1), 9 km (domain 2) and 3 km (domain 3) centered over central Nepal (26.34° N, 83.12° E) and the model is set to 40 vertical levels.

We also used a more realistic approach to investigate aerosol perturbation effects on precipitation response by implementing a prognostic CCN scenario. For this we modified the standard WRF model code to predict CCN as a function of time and position within the domain. We also used aerosol parameters that were derived from the Aerosol and Chemical Transport in Tropical Convection (ACTIVE) field campaign (Connolly et al., 2013) to initialize the model, the latter study using similar assumption for the ‘low’, ‘medium’ and ‘high’ aerosol scenarios as we did here.

To investigate the sensitivity of rainfall to aerosol and temperature perturbations we applied the range of initial conditions to three real rainfall events that occurred during different seasons of a one year period with varying rainfall intensity. This resulted in 45 WRF simulations in total. In the first case (Case I) the model was initialized at 00 UTC (11:45 Local) on 06 September 2007 and ran for 24 hours. This case was characterized by monsoonal flow from the Bay of Bengal, and may be referred to as a moderate intensity rain case. The second case (Case II) was initialized at 12 UTC (17:45 Local) on 25 September 2011 and ran
for 30 hours. This case was strongly influenced by moist monsoonal flow and representative of a high intensity rain event. In the final (Case III) the model was initialized at 00 UTC on 27 March 2011 and ran for 24 hours. This case is characteristics of a winter monsoon case influenced by a western disturbance, and was a low intensity rainfall event. In addition to the aerosol scenarios, the case-I was examined for the uniform temperature perturbation, random temperature perturbation and prognostic CCN scenarios, whereas the case studies II and III were investigated for the random temperature perturbation and the aerosol cases.

6. Results

Here we show the results of the sensitivity tests of rainfall to both aerosols and temperature perturbation using the three case studies. The effects of aerosols and temperature perturbation on the domain averaged accumulated rain (hereafter referred to as accumulated rain), domain and time averaged liquid water path (vertical sum of liquid water content in a column, hereafter LWP or liquid water path), domain and time averaged ice water path (vertical sum of ice water content in a column, hereafter IWP or ice water path) and domain and time averaged potential precipitable water (the depth of water in a column if all water vapour is condensed to liquid, hereafter PPW or potential precipitable water) are analyzed. Domain and time averaged cloud droplet effective radius and ice crystal effective radius (hereafter cloud droplet effective radius and ice effective radius respectively) are also evaluated. The Cloud Droplet Number Concentration (CDNC) along the south - north transect of the model domain as a function of updraft speed was computed. The effect of aerosol and temperature perturbations on droplet activation was elucidated by analyzing the 99th percentile of gridded column droplet number concentrations, which was represented by maximum simulated updraft speed. The evolution of hydrometeors was then analyzed using longitudinally averaged Hovmöller-like diagram and 2-D contour plots.
6.1 Activation of CCN and IN

6.1.1 Case study I

Here, we analyzed in detail, activation and evolution of cloud droplets ($R_{c,\text{eff}}$) and ice particles ($R_{i,\text{eff}}$) sizes in different aerosol and temperature perturbations cases. As expected, majority of our simulations show that greater aerosol concentrations give rise to clouds with smaller $R_{c,\text{eff}}$ (Table 5.1) and greater droplet number concentrations (Figure 5.2(a-c)). In contrast, no consistent effects of the aerosol perturbations were observed on the evolution of $R_{i,\text{eff}}$ (Table 5.2). Ice number concentration also remained unchanged (not shown), which is likely attributed to oversimplified parameterizations of ice nucleation processes as the chemical properties of aerosol are not taken into account (Muhlbauer et al., 2010). The temperature effects on cloud droplet activation were not large; however, they showed strong positive effects on ice nucleation processes, producing order of magnitude more ice crystals concentrations as the magnitude of the temperature perturbation increased (not shown). A good correlation between vertical velocity ($W_a$), domain and time averaged value calculated considering the grid points with positive vertical velocity (i.e. $W_a>0$), and CDNC (calculated for the corresponding grid points) was found (Figure 5.3). This is expected as the $W_a$ plays an important role to activate aerosol particles.
Figure 5.2: Correlation between Cloud Droplet Number Concentration (CDNC) and background aerosols concentration a) random temperature perturbation, b) uniform temperature perturbation, and c) prognostic CCN scenarios.
Figure 5.3: Correlation between domain and time averaged updraft speed (Wa) and CDNC in the uniform temperature perturbation along the north–south transect of the domain. Average CDNC is calculated for positive mean vertical velocity (Wa>0) in the corresponding grid box of the model. L, M and H in the legend stands for the ‘low’, ‘medium’, and ‘high’ aerosol cases respectively.

6.1.2 Case study II

The 99th percentile of CDNC reveals (not shown), as for the previous case, increases with increasing aerosol concentrations. The number of activated droplets shows positive correlation with aerosol number concentration irrespective of the temperature perturbation. Droplet activation is not significantly affected by the temperature perturbation as in the Case – I. A stronger positive correlation (c.f. case I) between CDNC and $W_a$ was found along south - north orientation of the domain (not shown). This may be expected as the clouds had the strongest, most organised updraft.
A decreased $R_{c,\text{eff}}$ was observed with an increase in aerosol concentration, however the effects of temperature was non-monotonic as shown in Table 5.1. Our simulations show that the $R_{i,\text{eff}}$ is slightly affected by aerosol number modification. In contrast, it is considerably affected by the temperature perturbations and the effects are inconsistent (Table 5.2).

**Table 5.1:** Inter-comparison of domain and time averaged cloud droplet effective radius ($\mu$m) for the set of simulations after 24 hours of simulations but after 30 hours of simulations for the case – II. Dashes indicate no simulations were performed for the case.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Case - I</th>
<th>Case - II</th>
<th>Case - III</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Low Medium High</td>
<td>Low Medium High</td>
<td>Low Medium High</td>
</tr>
<tr>
<td>Control (Tctrl)</td>
<td>14.85 14.80 14.77 14.84</td>
<td>14.82 14.78 15.21</td>
<td>15.21 15.21</td>
</tr>
<tr>
<td>Uniform temperature perturbation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tctrl+5</td>
<td>14.95 14.91 14.90</td>
<td>- - - - - -</td>
<td></td>
</tr>
<tr>
<td>Tctrl+10</td>
<td>14.93 14.89 14.86</td>
<td>- - - - - -</td>
<td></td>
</tr>
<tr>
<td>Random temperature perturbation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tctrl±0.5</td>
<td>14.84 14.81 14.76 14.85</td>
<td>14.82 14.79 15.21</td>
<td>15.21 15.21</td>
</tr>
<tr>
<td>Tctrl±2.5</td>
<td>14.85 14.82 14.78 14.85</td>
<td>14.82 14.80 15.21</td>
<td>15.21 15.21</td>
</tr>
<tr>
<td>Tctrl±5</td>
<td>14.81 14.75 14.72 14.82</td>
<td>14.78 14.77 15.21</td>
<td>15.21 15.21</td>
</tr>
<tr>
<td>Prognostic CCN</td>
<td>14.58 14.64 14.58</td>
<td>- - - - - -</td>
<td></td>
</tr>
</tbody>
</table>

### 6.1.3 Case study III

A positive correlation was observed between the 99th percentile of CDNC and aerosol number concentrations (not shown). Our simulation suggests that droplet activation is not affected by the temperature perturbations. Correlation between the CDNC and $W_a$ (not shown) indicated less number of activated droplets than the previous cases and low correlation (i.e. case study – I and case study – II). The $R_{c,\text{eff}}$ remained unchanged with increasing aerosol number concentration (Table 5.1) and also no significant effect was observed on evolution of $R_{i,\text{eff}}$. A non-linear effect of temperature was observed on growth of the $R_{i,\text{eff}}$ (Table 5.2) however, the effects were insignificant on the $R_{c,\text{eff}}$. 
Table 5.2: Same as Table 5.1 but for the ice effective radius (µm).

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Case - I Low</th>
<th>Medium</th>
<th>High</th>
<th>Case - II Low</th>
<th>Medium</th>
<th>High</th>
<th>Case - III Low</th>
<th>Medium</th>
<th>High</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control (Tctrl)</td>
<td>38.45</td>
<td>38.66</td>
<td>38.74</td>
<td>31.20</td>
<td>31.14</td>
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<td>26.43</td>
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<td>26.44</td>
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<tr>
<td>Uniform temperature perturbation</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tctrl+5</td>
<td>35.48</td>
<td>35.60</td>
<td>35.26</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Tctrl+10</td>
<td>37.12</td>
<td>37.36</td>
<td>36.89</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Random temperature perturbation</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tctrl±0.5</td>
<td>38.97</td>
<td>38.89</td>
<td>38.83</td>
<td>31.51</td>
<td>31.45</td>
<td>31.34</td>
<td>26.44</td>
<td>26.44</td>
<td>26.44</td>
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<tr>
<td>Tctrl±2.5</td>
<td>39.00</td>
<td>39.16</td>
<td>39.16</td>
<td>31.82</td>
<td>31.81</td>
<td>31.77</td>
<td>26.54</td>
<td>26.54</td>
<td>26.54</td>
</tr>
<tr>
<td>Tctrl±5</td>
<td>39.05</td>
<td>39.19</td>
<td>39.26</td>
<td>32.50</td>
<td>32.45</td>
<td>32.45</td>
<td>26.57</td>
<td>26.57</td>
<td>26.58</td>
</tr>
<tr>
<td>Prognostic CCN</td>
<td>29.87</td>
<td>29.36</td>
<td>29.22</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
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<td>-</td>
</tr>
</tbody>
</table>

6.2 Evolution of clouds and rain

6.2.1 Case study I

Figure 5.4 shows 2-D contour plots of 24 hour surface accumulated rain for the ‘low’ aerosol concentration, uniform temperature perturbation run. Shown in plot (a-c) are rain accumulations for the Tctrl, Tctrl+5 and Tctrl+10 runs.

In the ‘low’ aerosol case all simulations generally predict WNW – ESE oriented rainfall corridor except for the light intensity rain (<2.5 mm/hour) and few isolated patches. Our results show that the foothills of the Himalayas received more rain than in the high Himalayas (elevation >5000m) and ‘Teari’ areas (low land in the south, elevation <200m) during 24 hours. In the Tctrl run, the heavy rain area is confined across middle of central Nepal, mostly in the windward side of two well-known peaks (i.e. Champadevi and Phulchowki; elevation 2500 – 2700m) located SW – SE from the Kathmandu valley. These are well understood properties of orographic rain attributed to the complex Mountainous terrain (Barros and Lettenmaier, 1994) where warm and moist air is forced up to high elevation resulting in cooling and condensation / precipitation along the windward slope of the topographic barrier and subsidence in the leeward side evaporate the clouds.
Figure 5.4: 24 hours accumulated rainfall simulated for the ‘low’ aerosol concentration a) Control temperature (Tctrl), b) Tctrl+5°C, and c) Tctrl+10°C.

In the prognostic CCN control run (not shown), few patches of rain developed across central Nepal, mainly in the upwind of the Champadevi and Phulchowki peaks. A weak and unorganized rainfall corridor develops mostly along a west – east line, where the amount of rain is reduced considerably compared to the ‘low’ aerosol Tctrl run. However, the prognostic CCN run produced weaker rainfall across almost all of the domain from the low lands in the south to the high Himalayas in the north.

A drastic increase in the rainfall amount with an elongation of the rain corridor further northwest encroaching the higher Himalayas was observed in the Tctrl+5 and Tctrl+10 runs as shown in the figure. However, more pronounced effects are found in the latter. In contrast, in the random perturbation case, the organized rain belts gradually disintegrate into smaller patches, which generally move southward. The effect of temperature on rainfall was estimated to range from +37% to +93% in the ‘low’ aerosol and uniform temperature run depending on the simulation whereas it was -14% to +5% in the random perturbation case.
Hovmöller-like diagrams (Figure 5.5(a-c)), which show the evolution of hydrometeors over time along the south-north direction, revealed that a denser band of cloud developed across central Nepal (≈27°N – 28°N) early on in the control simulations and thin cloud evolved almost everywhere until the end of the simulation. In the uniform temperature perturbation case, the band of clouds rapidly developed as thick rain producing cloud over central Nepal, and gradually elongated towards the north with a rise in temperature, reaching almost to 29°N (South of the Tibetan Plateau) in the Tctrl+10 run. In contrast, thin clouds that developed later in the control simulation gradually diminish with increase in temperature and almost disappear in the Tctrl+10 run, which suggest a compensating process whereby a large local radiative effect is averaged out over the course of a day. In the random temperature perturbation, the dense cloud band enhanced to some extent in the Tctrl±0.5 run but weakens with perturbation magnitude and almost disappears in the Tctrl±5 run. However, there is little change in the thin cloud pattern and evolution.
Figure 5.5: Hovmöller-like diagrams for condensate simulated for the ‘low’ aerosol concentration a) Tctrl, b) Tctrl+5°C, and c) Tctrl+10°C.

Simulations of rainfall with the ‘medium’ aerosol are quite similar to the ‘low’ aerosol cases. In the Tctrl simulation the WNW – ESE rain corridor became more organized, expanded southward and slightly elongated along SE direction. The effects of temperature on rainfall were estimated to range from +35% to +90% in the uniform perturbation whereas it was approximated from -17% to
+2% in the random perturbation depending on the simulation. Hovmöller-like diagrams also revealed a very similar pattern of cloud development to the corresponding run in the ‘low’ aerosol case.

Figure 5.6: Sensitivity of microphysical variables to the uniform temperature perturbation a) liquid water path, b) ice water path, and c) potential precipitable water. Note that LWP includes liquid water from cloud drops and rain and IWP includes cloud ice, snow and graupel.
In the ‘high’ aerosol case, the simulated rain follows a similar pattern as described in the previous cases. The effects of temperature on rainfall was estimated to range from +31% to +92% in the uniform perturbation however it was found from -14% to +3% in the random perturbation depending on the simulation.

Figure 5.6 shows the sensitivity of (a) Liquid Water Path (LWP); (b) Ice water path (IWP) and (c) Potential precipitable water (PPW) to the aerosol and uniform temperature perturbations. The effects of aerosol perturbations are not significant on the simulated LWP, IWP and PPW; in contrast they are very sensitive to the temperature perturbations, as their magnitude increases with increases in temperature. Similar effects were observed in the random temperature perturbations but they were non-monotonic with increase in magnitude of the perturbation.

6.2.2 Case study II

In the ‘low’ aerosol case heavy rainfall occurred in the windward side of the foothills of the Himalayas forming a WNW – ESE rainfall corridor as in the case – I and light rain was scattered over the high Himalayas and ‘Terai’ areas. The effects of temperature on rainfall are estimated ranging from -11% to +3% in the ‘low’ aerosol case depending on the simulation. A well developed dense cloud is observed during the early hours of the simulations as indicated by the Hovmöller-like diagram (not shown), which reduces gradually with increase in magnitude of the random temperature perturbations.

In the ‘medium’ aerosol case spatial characteristics of accumulated rain follows quite similar patterns as in the ‘low’ aerosol case, occurring intense and concentrated episodes of rain over the foothills of the Himalayas and light rain across the ‘Terai’ and the high Himalayas. The effects of temperature on rainfall were estimated to range from -10% to +2% in the ‘medium’ aerosol case depending on the simulation.
In the ‘high’ aerosol case, the characteristics of rain are not significantly different from the ‘medium’ aerosol case; however, there are marginal changes in the spatial distribution of the heavy rain belt. The effects of temperature on rainfall were estimated to range from -11% to +3% in the ‘high’ aerosol case depending on the simulations.

The simulations show that the LWP, IWP and PPW are not very sensitive to the aerosol perturbations however show greater sensitivity to random temperature perturbations, but no consistent trends were found among the simulations.

6.2.3 Case study III

The rainfall distributions here in this case are completely different from the previous cases, as characteristic of the winter monsoon differ from the summer monsoon. In the ‘low’ aerosol case the high Himalayan region receives light rain and few patches of moderate rain. No rainfall corridor that formed in the previous cases was observed here. The light rain gradually moves southward as the magnitude of the random temperature perturbation increases producing several rain patches across the foothills and ‘Terai’ region in the Tctrl±5 run. The effects of temperature on rainfall are estimated to range from -1% to +22% in the ‘low’ aerosol case depending on the simulation. Patches of denser cloud are formed in some areas as revealed by the Hovmöller-like diagram (not shown) but scattered clouds remain for a sustained period of time, covering a wider area of the domain.

In the ‘medium’ aerosol case, the spatial distribution of accumulated rain is comparable to the ‘low’ aerosol case in the high Himalayas. However, development of rain is slightly reduced over the foothills of the Himalayas as the number concentration of aerosol increases and a significant effect can be observed in the Tctrl±2.5 and Tctrl±5 run. The effects of temperature on rainfall are estimated to range from 0% to +21% in the ‘medium’ aerosol case depending on the simulation.
In the ‘high’ aerosol case rainfall patterns over the high Himalayas are quite similar to the ‘low’ and ‘medium’ aerosol cases. The effects of temperature on rainfall are estimated to range from 0% to +21% in the ‘high’ aerosol case depending on the simulation. Hovmöller-like diagram of water condensate is also comparable to the ‘low’ and the ‘medium’ aerosol cases.

Our simulations show that the LWP, IWP and PPW are not significant to aerosol perturbations. The variables were slightly affected by the random temperature perturbations relative to the aerosol effects, but the effects are still insignificant, thereby not noticeable change in evolution of clouds.

6.3 Precipitation sensitivity to aerosol and temperature perturbation

6.3.1 Case study I

Figure 5.7(a-c) summarizes the key results for the effects of aerosol and temperature perturbation on rainfall in different perturbation scenarios. We observed that rainfall is slightly affected by aerosol number concentration, which is approximated from -3% to +2% and -2% to +4% with respect to the Tctrl run in the uniform and random temperature perturbation respectively. The effects of aerosol seem to be case dependent as rainfall can either increase or decrease with increasing aerosol number concentration. However, the simulations show that rainfall sensitivity is dominated by the temperature perturbation, which are expected to range from +31% to +93% and -5% to +17% in the uniform and random perturbation respectively. In the prognostic CCN control run the effects of aerosol on rainfall was estimated to range from -2.5% to +0.5%. A summary of accumulated rainfall from the sensitivity runs is presented in the Table 5.3.
Figure 5.7: Sensitivity of accumulated rain to aerosol and temperature perturbation a) random temperature perturbation, b) uniform temperature perturbation, and c) prognostic CCN scenarios.
Time series plots also indicate that rainfall is very sensitive to the temperature perturbation. In the random perturbation, the Tctrl run (not shown) indicates early development of rain which grows very rapidly, as a result, intense rainfall occurs in a short period of time. In contrast, although rainfall initiates nearly at approximately the same time, a more gradual evolution of rain was observed consequently produces less intense and prolonged rainfall in the other temperature perturbation cases (e.g. Tctrl±0.5, Tctrl±2.5 and Tctrl±5). This effect is more pronounced when the magnitude of perturbations is widened. In the uniform perturbation and ‘low’ aerosol case, as shown in the Figure 5.8, rainfall started at the same time in the Tctrl, Tctrl+5 and Tctrl+10 simulations however early evolution of rain followed by its rapid growth, is invigorated by enhanced moisture content, as was observed in the perturbed cases. These features of the rainfall are replicated in the ‘medium’ and ‘high’ aerosol scenarios and also observed in the prognostic CCN case (not shown).

**Figure 5.8**: Domain averaged time series rainfall for the uniform temperature perturbation ‘low’ aerosol concentration. Quite similar features of the rainfall were observed in the ‘medium’, and ‘high’ aerosol concentration cases.

Here we analyzed the structure of the cloud and precipitation regions (i.e. total condensate, hereafter referred to as condensate region) to elucidate their sensitivity to aerosols, which modulate their shape, size and spatial distribution.
Figure 5.9(a-c) shows a cross section of the condensate region along the point A and B in the Figure 5.4(a) just before onset of the rainfall for the ‘low’, ‘medium’ and ‘high’ aerosol case simulated for the Tctrl case. In the ‘low’ aerosol case a smaller condensate region both in horizontal and vertical extent with less hydrometeor content were observed. In the ‘medium’ and ‘high’ aerosol cases, the height of the condensate region and its horizontal extent was drastically increased, with more pronounced effects observed in the cold phase condensate that includes very large amounts of ice in the upper cloud regions. This may be attributed to increasing aerosol concentrations leading eventually to more riming and aggregation processes (Connolly et al., 2006). However this mechanism is still poorly understood.

Table 5.3: Inter-comparison of total rainfall (mm) for the set of simulations after 24 hours of simulations but after 30 hours of simulations for the case – II. Dashes indicate no simulations were performed for the case.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Case - I</th>
<th>Case - II</th>
<th>Case - III</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Low</td>
<td>Medium</td>
<td>High</td>
</tr>
<tr>
<td>Control (Tctrl)</td>
<td>34.5</td>
<td>35.0</td>
<td>35.2</td>
</tr>
<tr>
<td>Uniform temperature perturbation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tctrl+5</td>
<td>47.1</td>
<td>47.1</td>
<td>46.2</td>
</tr>
<tr>
<td>Tctrl+10</td>
<td>66.7</td>
<td>66.6</td>
<td>67.7</td>
</tr>
<tr>
<td>Random temperature perturbation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tctrl±0.5</td>
<td>36.3</td>
<td>35.6</td>
<td>36.1</td>
</tr>
<tr>
<td>Tctrl±2.5</td>
<td>32.4</td>
<td>32.4</td>
<td>31.8</td>
</tr>
<tr>
<td>Tctrl±5</td>
<td>29.6</td>
<td>29.0</td>
<td>30.2</td>
</tr>
<tr>
<td>Prognostic CCN</td>
<td>21.1</td>
<td>20.6</td>
<td>20.7</td>
</tr>
</tbody>
</table>

6.3.2 Case study II

Simulations of case study II suggest that rainfall is only slightly affected by the aerosol perturbations, as explained in case – I, and generally decreases with increasing aerosol concentrations (~2% to 0%). However, consistent trends were not found. In contrast, rainfall total was significantly affected by the random
temperature perturbations, which are estimated to range from -11% to +3%. Early evolution of rain and its rapid growth was observed in the time series analysis of rainfall in the Tctrl runs, which subsequently generate high intensity rainfall over a short period of time. A peak rainfall occurred after 5 hours of simulation in the Tctrl run, however, it was gradually shifted farther in the perturbed cases (e.g. Tctrl±0.5, Tctrl±2.5 and Tctrl±5) and the intensity was found to be significantly reduced with respect to the Tctrl.

![Graph a)](image1)

![Graph b)](image2)
Figure 5.9: Distribution of hydrometeors (g kg\(^{-1}\)) in the simulated condensate region (total condensate) for a) ‘low’, b) ‘medium’, and c) ‘high’ aerosol concentration for the control temperature simulation. These X-sections were taken at 4:45 GMT (10:30 am local time).

Analysis of individual cloud regions revealed they could extend to above 18 km, with freezing levels generally around 5 km. This indicates a dominance of cold phase cloud processes to the condensate. As in Case – I, our simulations indicated that the cold phase cloud region was significantly modified by the aerosol perturbations. A wider horizontal extent of the cold hydrometeor condensate region was observed in all the enhanced aerosol cases, again due to formation and transport of increasing amounts of liquid water to higher levels.

6.3.3 Case study III

Case study III showed only a weak rainfall sensitivity to the aerosol perturbations, which generally decreases with increasing aerosol concentration, ranging from -1% to 0%, however again no consistent trends were found. In contrast, we observed a strong sensitivity of rainfall to the random temperature perturbation, which ranged from -1% to +22% depending on the simulations.

Time series analysis (not shown) describes that the rain evolved almost at the same time in all the simulations, however gradual growth of the rain was observed in the perturbed simulations. We found that these effects are more
pronounced in the Tctrl±2.5 and Tctrl±5 runs and suggests that in general an increase in magnitude of the random perturbation delays precipitation onset.

Shape, size and spatial distribution of the condensate regions are not significantly affected by the aerosol perturbations in the case presented here. The ‘low’, ‘medium’ and ‘high’ aerosol cases produced a very similar structure in the condensate regions, which were mostly dominated by the cold phase cloud processes as it developed over the high Himalaya region. This could be attributed to a weak response of cold cloud ice processes to the aerosol perturbations. So the resulting effects of aerosols on cloud processing, spatial distribution and precipitation onset can be highly non-linear and case dependent.

7. Discussion and conclusions

A sensitivity study of aerosol - cloud - precipitation interactions was carried out in orographically generated mixed-phase clouds over the Himalayas.

We use a high resolution (3km x 3km) model to simulate the effects of aerosol and temperature perturbations on the evolution of clouds and rain. The sensitivity of rainfall to aerosol and temperature perturbations was analyzed for three observation case studies in different seasons with varying rainfall intensity contributions. We compared results from a control run to the perturbed temperature profile using both i) uniform, and ii) random perturbations, for three different cases, ‘low’, ‘medium’ and ‘high’ aerosol number concentrations. We also used a more realistic approach to investigate the effects of aerosol on rainfall with a prognostic CCN scenario, where the model explicitly predicts CCN as a function of time and location. In addition to rainfall and its onset time, the sensitivity of several microphysical parameters, including liquid water path, ice water path and potential precipitable water, to the perturbations was also evaluated.

The study shows the cloud droplets number concentrations are positively correlated to aerosol number concentrations, the effects are non-linear. Twomey
(1959) and Konwar et al. (2012) also suggest a non-linear increase in cloud droplet concentration due to an increase in CCN. Ice crystal number concentration is not much affected by the aerosol perturbations applied in our study, and this is attributed to the simplified parameterization schemes of ice nucleation processes in the model, where chemical properties of aerosol particles (Muhlbauer et al., 2010) are not considered. However, the ice crystal concentration is significantly enhanced by the temperature rise. Although the droplet size is decreased with increasing aerosol number concentration, the size of the cloud ice is generally increased. It may be qualitatively argued that increasing the number of smaller droplets results in the ice particles at higher altitude having liquid water for riming and this could have contributed to increase size of the ice particles. However, this feature is less robust as it is not supported by all scenarios. Activation of cloud droplets increases with increases in updraft speed as expected.

It is found that amount of rainfall is slightly sensitive (-3% to +4%) to aerosol number concentration and the effects are non-linear. With respect to these sensitivities it can be discussed that the contribution of ice phase is dominant in the precipitation processes as the LWP/IWP ratio ranges from 0.5 – 0.9 in the random and 0.5 – 0.7 in the uniform temperature perturbation, which could compensate or even outweigh the suppressed warm rain via melting of ice hydrometeors at aloft (Lee and Feingold, 2010). A poor sensitivity of ice crystal concentration to the aerosol perturbation found is similar to the reported sensitivity of precipitation to aerosol concentrations in simulations of orographic clouds in the Alps or Rocky Mountains (Muhlbauer et al., 2010).

Early evolution of rain is observed in a warming climate but it is sustained for a very short period of time. We found that magnitudes of such effects are more pronounced with rising temperatures, as +31% to +93% more rain was produced in the uniform temperature perturbation cases. These effects on the rainfall simulations are attributed to the increase in atmospheric moisture caused by the increased temperature, which suggests that at constant relative humidity moisture content increases at about 6%/ K (Soden et al., 2005). Thus, Indian monsoon rainfall over the Himalayas (Turner and Annamalai, 2012) and also tropical precipitation (Allan and Soden, 2008) are affected by global warming.
Our simulations indicate that the windward side of the foothills across central Nepal may receive more rain during the summer monsoon period and a precipitation corridor evolves along WNW – ESE orientation (i.e. parallel to the Himalayas). During the summer monsoon, moist air from the Bay of Bengal is transported towards the Himalayas and encounters the giant barrier which in turn develops orographic cloud through terrain-forced convection and associated precipitation. However, this mechanism is not observed during the winter monsoon season as suggested by our simulations.

Analysis of total condensate reveals a change in aerosol concentration likely to modify the horizontal and vertical distribution of hydrometeors. Smaller size of the condensate region with lesser amount of hydrometeors content in the ‘low’ aerosol case is drastically increased in the ‘medium’ and ‘high’ aerosol cases, more pronounced effects are found in the ice phase clouds, where it is argued that invigoration of riming and aggregation processes become efficient through the abundant supply of super-cooled liquid water at the higher altitude. Hence our simulations suggest aerosol perturbations can modify the shape, size and spatial distribution of individual cloud regions and their precipitation.

8. Limitations of the study

The dominance of the ice process in the Himalayan clouds highlights the importance of the need for further work to assess the sensitivity of the Himalayan clouds to IN perturbations. In the study conducted here it was assumed that both the spatial variation in IN concentrations and their relative contribution to the total aerosol number will remain the same in all cases, which may not be true. A strong sensitivity of Arctic mixed-phase clouds to IN was observed (Morrison et al., 2005b), which further support the hypothesis that the Himalayan clouds are sensitive to IN. Our fundamental understanding of the ice phase processes is improved by recently developed ice nucleation scheme, however further study is needed for accurate model representations (Connolly et al., 2013).
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Reference


Aerosol mass and black carbon concentrations, a two year record at NCO-P (5079m, Southern Himalayas), Atmos. Chem. Phys., 10, 8551-8562, doi:10.5194/acp-10-8551-2010, 2010.


CHAPTER 6: General Conclusion

This chapter summarizes the findings of this thesis. Detailed conclusions have been described at the end of each paper so only key findings are mentioned here. The thesis is comprised of three journal papers, one of the manuscripts has already been submitted and the remaining two papers are being submitted for publications, and are summarized in the subsequent sections. Furthermore, contributions of this thesis to science and general public, limitations and future works are also described.

6.1 Summary

This thesis has attempted to analyze the sensitivity of microphysical schemes to changes in temperature and aerosol concentration in order to improve our understanding of precipitation patterns over the Himalayan region of central Nepal. The main motivation of this study was to explore level of sophistication required from a microphysics scheme to represent aerosol– precipitation interactions in the Numerical Weather Prediction (NWP) model when it is applied to the topographically complex area. The study was carried out using the cloud-resolving Weather Research and Forecasting (WRF) model supported by limited field observation.

The first paper (chapter 3) summarizes status quo information of surface meteorology and aerosol size distribution at a mid-hill region of central Nepal (Nagarkot, altitude ~1900 m), which provides background aerosol information to use for the modelling study in this thesis. The observed data is also important for validation of the model output so this paper is a foundation of the rest of the thesis. The aerosol particles were measured in 31 different size channels ranging from 0.25 µm to 32 µm, with one minute temporal resolution, which was lasted about a year from March to December, 2011 that made possible to analyze diurnal and seasonal variation of particles size distributions and its interaction with meteorology. However, one of the criticisms of this research is that the particle
measurement at Nagarkot does not include the particles with diameter less than 0.25 µm where most of the climatologically active particles are present. We summarize our main findings as below:

- A strong seasonal variation of aerosol total particles was observed with maximum concentration occurred during the post-monsoon season (919±694 cm\(^{-3}\)) followed by the pre-monsoon (746±561 cm\(^{-3}\)) and the monsoon seasons (374±361 cm\(^{-3}\)). A similar seasonal variation of aerosol number concentration was observed in the Tibetan Plateau region (altitude ~4180 m), however the maximum concentration was observed in the pre-monsoon season which was associated with the dust storm in the Northern China (Xu et al., 2013). The observed number concentration of aerosol at the Tibetan Plateau region was significantly lower than our observation at the mid-hill region. Furthermore, Monkkonen et al. (2003) also found maximum particle concentration in March in New Delhi, India but the number concentrations were several order of magnitude more than our observation, which may indicates an altitudinal variation of aerosol number concentration.

- Diurnal variations of particle concentrations showed a unique twin peaks occurring one in the morning and the other in the evening. These diurnal trends were attributed to the planetary boundary layer (PBL) activities and thermal wind circulations. A strong influence of PBL and thermal wind circulations was also reported on the diurnal cycle of aerosol particles in the other mountainous terrain, for example in the Tibetan Plateau region (Xu et al., 2013), Whistler mountain region, Canada (Gallagher et al., 2011), Pico Espejo, Venezuela (Schmeissner et al., 2011) and Jungfraujoch (Lugauer et al., 2000). Our results indicated that these peaks were generally caused by enhancement of coarse size particles, which suggested that increased local activities might have played a crucial role.

The second paper (chapter 4) compares four different bulk microphysical parameterization schemes (WSM6, WDM6, Lin and Morrison schemes) available
within the WRF model. The schemes, which generally increase in complexity in terms of microphysical processes, range from simple single moment microphysics scheme to sophisticated double moment scheme. The idea was to analyze the sensitivity of the microphysics schemes and statistically compare the controlled model outputs with observation. A convective storm that evolved over central Nepal during the pre-monsoon season (March – May) was simulated as a case study. The key findings are summarized as below:

- The performances of the four microphysical parameterization schemes under controlled conditions were statistically evaluated with the observed meteorological variables at Nagarkot Nepal. Among the different microphysical schemes used in the simulations, the Morrison double moment scheme which predicts number concentration and mixing ratio of the hydrometeors performed best. It did not show any statistically significant difference compared to the observed rainfall data at 80% confidence interval using a Chi-squared goodness of best fit test. Jankov et al. (2011) found that the Morrison double moment scheme performed better than the Lin scheme in the performance evaluation of the WRF microphysics schemes. Likewise, Molthan and Colle (2012) indicated that the Morrison double moment schemes performed better than the single moment schemes in the WRF microphysics inter-comparison study.

- A strong sensitivity of microphysical parameterization schemes was occurred to the ice phase hydrometeors and cloud cover. The Morrison scheme produced more upper level clouds and snow mixing ratio as compared to the other schemes; however graupel content in the scheme was the lowest. These effects were attributed to the different assumptions made in the parameterizations of cloud processes; for example, the schemes use different threshold size to form snow and graupel. Furthermore, not all the schemes include the secondary ice formation processes and also they have different assumptions while estimating the intercept parameter of the hydrometeors. Moreover, the sensitivity may also have been caused by moment of distribution of the hydrometeors as reported by Shipway and Hill (2012) and Molthan and Colle (2012) in the
inter-comparison of microphysics parameterization in the cloud-resolving model where they found that the double moment scheme performed better than the single moment scheme.

The third paper (chapter 5) investigates sensitivity of Indian monsoon precipitation to aerosol and temperature perturbations. The aerosol perturbations used in the simulations were based on our field observation described in the chapter 3 (1st paper) and also from the published literatures, whereas temperature perturbations were estimated from the IPCC SRES (Special Report on Emission Scenarios) scenarios. We used the Morrison double moment bulk microphysics scheme based on the performance evaluation of the microphysics schemes described in the Chapter 4 (2nd paper). Herein we aim to investigate how aerosol and temperature perturbations may affect cloud microphysical processes and thereby precipitation evolution in the cloud-resolving simulations over the complex Himalayan terrain. The key findings are summarized as follows:

- The effects of aerosol on the precipitation were insignificant and non-linear (Connolly et al., 2013). Increased aerosol number concentrations thereby CCN gave rise to cloud droplets number concentrations which in turn reduced the droplets size (Xie et al., 2013). However, ice crystal concentrations were not significantly affected by the aerosol perturbations (Lim and Hong, 2012). The weak sensitivity of aerosol to precipitation can be described that there was little contributions from warm-rain process to precipitation and thus the processes are dominated by cold phase clouds processes, which is poorly represented in the current climate and weather models. Inclusion of chemical properties of aerosol (Muhlbauer et al., 2010) and sophisticated ice nucleation processes involving habits of ice crystals in the simulations could provide better results. Furthermore, sensitivity analysis of precipitation to Ice Nuclei (IN) is also recommended.

- The effects of temperature perturbations on the precipitation were more than the aerosol effects. An early evolution of rain followed by its rapid growth and instant dissipation was observed in a warming climate. These
effects on the rainfall were attributed to increase in atmospheric moisture caused by the increased temperature. Furthermore, increase in temperature may increase updraft speed due to latent heat release that could invigorate the storm development.

- A considerable effect of aerosol was observed in the individual cloud regions and the effects were more pronounced in the ice phase clouds. Increased horizontal and vertical extent of clouds was observed with increase in aerosol concentrations. This mechanism may be attributed to invigoration of riming and aggregation processes through the abundant supply of super-cooled liquid water at the higher altitude. This argument is also supported by increased size of ice crystals, although it is less robust.

6.2 Contributions to Science and general public

Chapter 3 describes overview of current air pollution situations and its possible dispersion mechanisms of the Kathmandu valley. In addition to the diurnal circulations, it also provides general understanding of seasonal cycles of the pollutants within the valley. This thesis also made significant contributions to science. The pollution and meteorology data that we collected during our field observation is available for future researchers who are interested in air quality research across the region. The pollutants ventilation from the valley generally take place from late morning to early afternoon, so emission related activities such as operation of brick kiln industries, garbage incineration, movements of heavy trucks need to be restricted during this period to reduce pollutants residence time within the valley. Furthermore, location of industries can also be limited to eastern passes (e.g. the Nala and Sanga) area, as the strong westerly winds swipes out the valley’s pollutants through the passes. It is generally considered in Nepal that an early morning walk is good for health but it would probably expose the valley resident to the more polluted air in the morning.

Another key achievement of this thesis was a successful application of the high resolution Weather Research and Forecasting (WRF) model in the complex
Himalayan terrain, which captured very detailed features of the clouds and precipitation development, although further works are needed to accurately represent the processes. As the PhD is a collaborative work between a local research institute in Nepal and the University of Manchester, which has established a good scientific relationships between two countries, in turn open doors for further collaborative works in the future.

6.3 Limitation

A number of shortcomings have been realized, which has limited the scope of this thesis. These were lack of local experts to handle the equipments, problems in the equipments during the field work, power outage and unavailability of resources for the field observations.

Field observation presented here in this thesis entirely based on single measurement station that limits our understanding of aerosol size distribution and dispersion mechanisms and in turn the model simulations. Although the original plan was to deploy two sets of equipments, one in the mid-hill peri-urbane area (Nagarkot) and another in the high altitude remote location (Lukla), limited by non-availability of local experts to operate the equipments in Lukla and also shortage of funds associated with battery back-up system and instrument running cost.

Instrumental problems were also observed during the field experiment. From the very beginning of the experiment humidity sensor of the GRIMM particle counter did not work for some reason; however we recorded it from the Vaisala weather sensor. Although we provided battery backup system for the power outage period, this system was also not fully reliable; consequently we have higher frequency of data gaps. Due to water leakage or some other reasons the particle counter also did not work after middle of November, 2011 at Nagarkot site. In the beginning of 2012, we shifted the equipments at the city center with an aim to compare aerosol size distribution between city center and peri-urbane site.
However due to lack of local exports we failed to set up the equipments properly as a result the whole system did not work at all.

Vertical profile of the atmosphere is an important source of information to validate the model, however non-availability of Helium gas in Nepal prevent us from lunching the weather balloons, in turn we are unable to compare observed vertical profiles of the atmosphere to the WRF simulated profiles.

Besides these problems I faced a number of family problems such as serious health condition of our son and my wife, which significantly interrupted progress of the work and the situation was further worsen by loss of our family member in the final year of this PhD.

### 6.4 Future work

As described, this thesis has improved our understanding of seasonal and diurnal variations of aerosol particles size distribution and its interactions with meteorology over the complex Himalayan terrain through the field observation and cloud-resolving model simulations. We also described possible pollution dispersion mechanism of the Kathmandu valley, Nepal. Since our simulations showed that ice phase cloud process plays a crucial role to buffering the sensitivity of rainfall to aerosol perturbations. It is very likely that oversimplified parameterization of ice nucleation processes used in the model may have played the role. So, model simulations with aerosol chemical characteristics could improve our understanding of aerosol – precipitation interactions over the region. Furthermore, sensitivity of precipitation to IN is also important in understanding clouds processes (Morrison et al., 2005).

Precipitation formation processes over the South Asian region is dominated by convective phenomenon, which could transport moisture and chemical tracers higher up, even in the lower stratosphere (Oltmans and Hofmann, 1995), could have serious implications on the global climate. Furthermore, deep convection over the foothills of the Himalayas during boreal summer season occurs as frequent as in the Tropical region due to warm and moist monsoonal
flow and orographic lifting (Luo et al., 2011). The detailed features of the convection and associated precipitation formation mechanisms are poorly represented in the current climate models, particularly the cold phase cloud processes are very limited. Recent studies reveal that Satellite derived data could improve our understanding of ice phase processes and thereby current climate and weather models (Li et al., 2007).

We attempted to evaluate the WRF model using the Cloudsat/CALIPSO satellite data, which provides detailed features of cold and warm phase cloud hydrometeors, with an aim to improve parameterizations of cloud microphysical processes in the cloud-resolving model however it was not possible within the time frame of this PhD. Some figures have been included here that intrigues for further research. Figure 6.1(a-b) shows radar reflectivity observed by the Cloudsat satellite over central and eastern Nepal during day and night time overpass respectively, which shows strong nocturnal convection over the foothills of the Himalayas. This mechanism has been reported in some previous field observations but not well documented yet. So, rigorous modelling study supported by field observations and satellite retrievals would be necessary to understand the processes, particularly focusing the following questions:

- Does this mechanism really exist or just a coincidence or artefact? Very few field studies focusing on a small scale river basin level have been conducted to examine this mechanism. Furthermore, no modelling studies have been carried out to investigate this process. Hence, realism of this hypothesis needs to be tested and verified.

- What are the driving factors that cause nocturnal convection over the Himalayan foothills? A few number of field measurements have been conducted across central and eastern Nepal, which indicated that interactions of weak up-valley flow with the katabatic wind were responsible for this mechanism. However, more recent study pointed out that this mechanism was dominant during the monsoon season and associated with the interactions between moist south-westerly monsoon flow and diurnal circulation of mountain-valley wind.
• What is the difference between rainfall characteristics generated by diurnal and nocturnal convections? Previous studies pointed out that ridges and peaks receive diurnal maximum precipitation in contrast, valley receives nocturnal precipitation, which is attributed to formation of convective clouds during the day and night time.

• Is this mechanism limited to particular seasons or occur throughout the year? Previous field observations were carried out during monsoon season so conclusions entirely based on limited summer time observations.

Figure 6.1: Vertical cross sections of radar reflectivity (colour shedding) measured by the Cloudsat satellite a) day time pass over central Nepal on 07 July 2009, and b) night time pass over eastern Nepal on 21 July 2009. Black and red dots represent locations of cloud-layer top and bottom respectively. Blue lines top of each panels indicate deep convective cores. Black shedding represents elevation of the land surface.
6.5 Reference


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