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Key Points:

- Synoptic-scale differences lead to differences in mountain snowfall
- Snowfall rate in mountains is determined by cross-barrier moisture flux
- Future climate change in the region will affect snowfall during these events

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WRF simulations of two extreme snowfall events associated with contrasting extratropical cyclones over the western and central Himalaya

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Abstract Two extreme snowfall events associated with extratropical cyclones, one interacting with the western and one with the central Himalaya, are simulated with the Weather Research and Forecasting (WRF) model over 8 days. One event in January 1999 was driven by a longwave trough over west Asia, with the cyclone becoming terrain-locked in the western Himalayan notch. Another event in March 2006 was driven by a trough further south and east, facilitating the passage of two successive cyclones along the entire Himalayan ridge, drawing moisture from warm tropical waters. These flow patterns are typical for extreme winter precipitation in the western and central Himalaya, respectively, but were amplified in these two cases. In the WRF simulations, snowfall is confined to the western Himalaya in the January simulation, while a near-continuous band of accumulated snowfall along the Himalayan ridge forms in the March simulation. Precipitation rate throughout both simulations is largely determined by cross-barrier moisture flux, which is generally greatest wherever the cyclonic winds converge on the mountains at each time. However, the March 2006 simulation evolves in an environment with greater convective instability upwind of and moisture transport toward the mountains than in the January 1999 event. Hence, greater precipitation rates and more solid snowbands are generated in the March than in the January simulation. However, due to the terrain-locking of the cyclone in the January event, individual locations receive more persistent snowfall, so that the greatest 8 day accumulations are similar between the two events, although these accumulations are more widespread in the March event.

1. Introduction

The precipitation associated with extratropical cyclones may be enhanced and redistributed when the cyclone encounters orography [e.g., Braun *et al.*, 1997; Dettinger *et al.*, 2004; Medina *et al.*, 2007; Rasmussen *et al.*, 2011; Viale *et al.*, 2013]. In the Himalayas, winter extreme precipitation events typically occur in conjunction with extratropical cyclones, normally between December and March, termed westerly disturbances, with moisture gained from the Mediterranean, Arabian, or Caspian seas [e.g., Lang and Barros 2004; Barlow *et al.*, 2005; Hatwar *et al.*, 2005; Barros *et al.*, 2006; Dimri 2006; Cannon *et al.*, 2015]. There are typically just between two and four strong westerly disturbances per winter, providing the vast majority of the total winter precipitation to the mountains. Therefore, whether an individual event affects a given location in the mountains may significantly change water availability within that drainage basin. The Himalayan drainage basins provide water for millions of people in South and central Asia. For this reason, an improved understanding of mesoscale processes within extratropical cyclones that determine how much precipitation falls into the different drainage basins is paramount to predicting moisture availability in the region.

In addition to the lift provided by the mountains, particular geographical features may significantly alter the winds associated with extratropical cyclones and, thus, the development of the storms. There are two notches in the High-Asia Mountains (Figure 1a) in which a low-lying area is surrounded by tall mountains. Lang and Barros [2004] investigated severe snowstorms associated with upper level disturbances passing over the western notch between the western Himalaya and Hindu Kush Mountains. They found that the location of a disturbance on the subtropical jet, relative to the Tibetan Plateau, may determine whether a cyclone becomes terrain-locked, meaning that the cyclone remains quasi-stationary in the area enclosed by the orographic barrier. Therefore, the location of the jet, relative to the topography, is crucial to the mesoscale distribution of winter precipitation in the mountains.

Figure 1. The domains used to simulate both the January 1999 and March 2006 events. (a) The outer domain at 20 km grid spacing, showing the location of the inner domain. (b) The inner domain at 6.7 km grid spacing. Plotted in both panels is elevation (greys, km), showing the resolution of the terrain at the two grid spacings. In Figure 1b, the boundaries of all drainage basins draining across the Himalayan range are also drawn, plotted from a shapefile interpolated onto the inner domain's grid spacing. The "Western notch" and "Eastern notch" annotations in Figure 1a are for reference in the text.

In southern Asia, the subtropical jet has climatologically shifted northward and intensified since the 1980s [e.g., Cannon *et al.*, 2015]. Acting as a baroclinic-wave guide, the jet's northward shift has restricted cyclones from passing south of the Himalayan ridge, so that the central Himalaya has been affected by fewer cyclones. The Karakoram and western Himalaya, however, have been increasingly affected, as more cyclones are prone to being trapped in the notch [Cannon *et al.*, 2015]. These trends have contributed to the advance of many glaciers in the Karakoram and western Himalaya, and the retreat of glaciers in the central and eastern Himalaya at greater rates than almost any others on Earth [e.g., Bookhagen and Burbank 2010; Scherler *et al.*, 2011; Bolch *et al.*, 2012]. The current study is not concerned with interannual and decadal trends in these storms but seeks to describe the response of the mountains to given synoptic-scale conditions. The above discussion illustrates how the different areas within the Himalayas may be profoundly affected by changes to extreme precipitation events, due to subtle differences in the synoptic-scale flow pattern.

An observational approach is of limited use for investigating these events because there is no precipitation radar and rain-gauge data in the Himalayas (normally of snowfall) are sparse and unreliable [Qin *et al.*, 2009], particularly during extreme conditions. Gridded data sets of estimated precipitation, the

Tropical Rainfall Measuring Mission (TRMM) and Asian Precipitation Highly-Resolved Observational Data Integration Towards Evaluation of Water Resources (APHRODITE), are available in the region and TRMM is utilized in this study. However, these products have known deficiencies (see section 4) and, at 0.25° lat–lon grid spacing, fail to capture mesoscale features and precipitation structures arising from the complexities of the mountains. Global reanalyses are also too coarse for such purposes. Therefore, mesoscale model simulations must be performed to investigate the spatial and temporal distribution of this extreme snowfall at high elevations.

Previous studies have found that mesoscale simulations over the High-Asia mountains may agree well with the limited observations available [e.g., Azadi *et al.*, 2001; Lang and Barros 2004; Barros *et al.*, 2006]. In particular, the Weather Research and Forecasting (WRF) model has been shown to be capable of accurately simulating known patterns of extreme precipitation in the High-Asia mountains [Sato *et al.*, 2008; Li *et al.*, 2009; Maussion *et al.*, 2011, 2014; Sato 2013] and in other high mountains around the world [e.g., Rasmussen *et al.*, 2011; Jiménez *et al.*, 2013; Viale *et al.*, 2013]. Maussion *et al.* [2011] performed a month-long WRF simulation over the Tibetan Plateau, with a particular focus on 1 week in which heavy

precipitation was observed. They found WRF to be accurate in simulating rain and snowfall over a month, according to rain gauges and show a marked improvement from TRMM in doing so. They also found the simulations to be fairly insensitive to the choice of physical parameterizations employed.

The greatest endorsement of mesoscale simulations in the High-Asia Mountains comes from the study of *Maussion et al.* [2014] who performed a long sequence of daily simulations with WRF. These simulations produced the High Asia Reanalysis (HAR), a model data set at 10 km grid spacing, spanning 2000–2011, serving as a dynamical downscaling of coarser reanalyses to showcase the intraseasonal, intraannual, and interannual variability of mesoscale features in the region over the given time period. *Maussion et al.* [2014] found the HAR to recreate previously reported patterns of spatial and temporal variability of precipitation in the region. The HAR indicates that mesoscale models, and in particular WRF, are capable of accurately simulating the interaction of various types of weather systems with these mountains to generate precipitation features, at least on the scale resolved by 10 km grid spacing.

However, the above study was limited to a climatology of the region and did not examine any specific extreme events, which, as argued above, are critical to the moisture supply to different areas within the region. Furthermore, the 10 km grid spacing employed in that study may not capture certain flow features generated by complex terrain. For example, *Rasmussen et al.* [2011] found that only with grid spacing of 6 km or below were WRF simulations over the Colorado Headwaters regions accurate in simulating observed spatial patterns and precipitation amounts. In coarser simulations in that study, the snowfall was evenly distributed between valleys and elevated areas, due to the weak and broad vertical motion arising from shallow terrain. In higher-resolution simulations, the snowfall was more concentrated into the mountain peaks, more closely resembling observations, due to the strong and narrower updrafts arising from the steeper terrain.

Lang and Barros [2004] found that significant winter precipitation in the Himalayas only occurs when the large-scale flow evolves to a favorable geometry with respect to the mountains and hypothesized that wind speed and moisture flux in the cross-barrier direction should determine snowfall amount. This relationship has been found to hold in other mountainous regions [e.g., *Browning et al.*, 1974; *Neiman et al.*, 2009]. Therefore, the cyclones with the strongest cross-barrier winds and those that evolve in an environment with the greatest moisture supplies should lead to the greatest snowfall. To this effect, *Barros et al.* [2006] simulated an event in February 2000 in which an extratropical cyclone passed east along the Himalayan ridge, avoiding terrain-locking. As the southerly winds on the eastern flank of the cyclone flowed against the central Himalaya, a wave-like perturbation in vertical velocity was generated across the mountains, leading to along-barrier precipitation bands.

All of the above studies demonstrate that a great number of synoptic and mesoscale factors are relevant in determining the distribution and intensity of precipitation when extratropical cyclones interact with high mountains, and that mesoscale models can describe the interplay between these factors. Despite the modeling studies discussed above, there has not been a sensitivity study of extreme precipitation associated with extratropical cyclones in the Himalayas to synoptic and mesoscale flow features. To this end, this paper uses WRF to simulate two cases of extratropical cyclones in the Himalayas, primarily affecting the western and central Himalaya, respectively, in which the influences on precipitation in the mountains were strongly contrasting from one another. Because this paper is primarily interested in extreme precipitation in the Himalayas, the focus of the simulations is on precipitation at high elevations, i.e., snowfall. Thus, relatively little attention is paid in this study to simulated rainfall at the foot of the mountains that does not impact mountain hydrology. By comparing the two events, in terms of the above synoptic and mesoscale factors, and in terms of the spatial and temporal distribution of simulated snowfall, we seek to gain insight into the most important influences on snowfall during westerly disturbances, at least in these two cases.

The remainder of this article is organized as follows. Section 2 presents the synoptic backgrounds of the two cases studied and outlines the crucial differences that lead to variability in simulated snowfall presented later in the paper. Section 3 details the mesoscale model used and its configuration. Section 4 describes the gridded-precipitation estimates to which the simulated precipitation is to be compared. Section 5 provides an overview of the snowfall evolution in the simulations of each event, before section 6 shows the spatial distribution of precipitation accumulated over the simulations. Section 7 then examines more closely how the model simulates the interaction between the synoptic-scale forcing and the mountains to generate the snowfall presented in the previous sections. Finally, section 8 concludes this article.

2. Synoptic-Scale Background and Identification of Events

This paper focuses on two events in which extratropical cyclones led to extreme snowfall in the Himalayas: 3–11 January 1999 (hereafter, J99) and 9–17 March 2006 (hereafter, M06). The events were selected to exhibit winter storms bringing extreme precipitation to the western and central Himalaya, respectively. However, as will be shown, storms affecting the central Himalaya also tend to affect the western Himalaya before they move east toward the central Himalaya.

To identify the events, similar to Cannon *et al.* [2015], TRMM 3B42 precipitation estimates over the Himalayas were taken everyday in December to March from 1998 to 2013. The total precipitation over the western Himalaya and central Himalaya was calculated each day (see Cannon *et al.* [2015, Figure 2a] for the precise areas over which precipitation was summed for each of these regions). All independent dates in the 95th percentile of total precipitation over the western Himalaya were used to construct composites using the Climate Forecast System Reanalyses (CFSR) [Saha *et al.*, 2010] from the National Oceanic and Atmospheric Administration (NOAA)'s National Centers of Environmental Prediction (NCEP) at 0.5° (pressure levels) and 0.31° (surface variables) lat–long grid spacing. For two dates to be considered independent, they had to be at least 3 days apart, so that only the day with the greatest total precipitation was retained. These composites are shown, together with the corresponding TRMM composite, in the left-hand panels of Figure 2. As documented in Cannon *et al.* [2015], these composites show that extreme winter precipitation falls in the western Himalaya when a longwave trough in the subtropical jet just west of the Himalayas (Figure 2a) leads to southwesterly moisture advection from the Arabian Sea normal to the western Himalayan mountain front (Figure 2c). One date constituting these composites, 6 January 1999, is an extreme example of these characteristics. A greater-amplified trough (cf. Figures 2a and 2b) led to stronger moisture advection from further south (cf. Figures 2c and 2d) than in the composites. Consequently, TRMM precipitation for 6 January 1999 is similarly distributed but much greater than in the composite (cf. Figures 2e and 2f).

The same analysis was performed for dates bringing 95th percentile precipitation to the central Himalaya (Figure 3). Extreme winter precipitation falls in the central Himalaya when a trough further south and east than in the western Himalaya composite (cf. Figures 2a and 3a) leads to moisture advection from the Arabian Sea normal to the central Himalayan mountain front (Figure 3c). Moisture is also advected toward the mountains anticyclonically from the Bay of Bengal. The precipitation extends further along the Himalayan front than in the western Himalaya composite (cf. Figures 2e and 3e) because cyclones affecting the western Himalaya become trapped in the western Himalayan notch, as discussed in section 1. One date constituting these composites, 10 March 2006, is as an extreme example, enhancing all of these characteristics (cf. left and right columns of Figure 3). In particular, a strongly southward-displaced trough above western India (Figure 3b) led to particularly strong southerly moisture transport from the Arabian Sea and Bay of Bengal all the way through India toward the central Himalaya (Figure 3d) and hence a band of intense precipitation along the central Himalaya (Figure 3f).

Thus, 6 January 1999 and 10 March 2006 were selected as examples of extreme precipitation in the western and central Himalaya, respectively, when the synoptic-scale pattern was morphologically typical for extreme precipitation in the given area, but when this pattern was strongly amplified, even amongst 95th percentile dates. The contrast between western and central Himalaya extreme-precipitation dates shown in the composites imply a stronger contrast between these two dates. Comparison of the two dates in this paper will illustrate how cyclones evolving in contrasting environmental conditions interact with topographic features and contribute to the spatial variability of excessive precipitation across the Himalayas.

The full events, J99 and M06, in which these two dates were embedded were identified by examining TRMM data and identifying the range of dates over which extreme precipitation occurred somewhere along the Himalayan range. However, the M06 event consists of more days of extreme precipitation than J99, as will be described, and so J99 was extended in order to compare the same number of days between simulations.

The J99 event occurred during a moderate La Niña episode, while M06 occurred during a weak La Niña episode (<http://ggweather.com/enso/oni.htm>). Syed *et al.* [2006] and Yadav *et al.* [2009] both found that wintertime precipitation in the High-Asia Mountains is enhanced during El Niño episodes. This enhancement occurs via the development of a trough over southwestern Asia, intensifying westerly disturbances and steering them toward the mountains. However, as discussed above, according to TRMM,

Figure 2. Comparison between composite of extreme-precipitation days in western Himalaya and 6 January 1999 of CFSR (top two rows) and TRMM 3B42 (bottom row). (a,b) 1000–500 hPa thickness (colors, dam) and 300 hPa geopotential height (black contours every 50 m, labeled every 100 m). (c,d) Precipitable water (blues, mm) and column-integrated moisture flux (vectors, $\text{kg m}^{-1} \text{s}^{-1}$). (e,f) Daily accumulated precipitation (mm, colors). Also plotted in all panels in bold is the 3 km elevation contour, indicating the location of the High-Asia mountains. Composites constructed as detailed in section 2. Right-hand panels show the 00 UTC 6 January reanalysis (Figures 2b and 2d) and the 00 UTC 6 January to 00 UTC 7 January accumulated precipitation (Figure 2f).

Figure 3. As Figure 2, but comparing composite of extreme-precipitation days in central Himalaya with 10 March 2006.

6 January 1999 and 10 March 2006, occurring in La Niña conditions, were amongst the most extreme days of winter precipitation from 1998 to 2013 in the western and central Himalaya, respectively. According to TRMM, the accumulations from the two events over 8 days constituted the majority of their respective winter accumulations (1 December to 31 March) over large areas, particularly M06, which was the only event to deposit major snowfall in the central Himalaya that winter (not shown). However, according to TRMM, both these winters were drier than average in southern Asia, possibly due to the La Niña episodes throughout both winters. Therefore, La Niña episodes may reduce total winter precipitation in the Himalayas but still hold potential for individual intense storms to generate some of the largest precipitation accumulations.

Figure 4. Daily infrared imagery from Meteosat over southern Asia during the two events at the given times and dates.

The evolution of the two events is illustrated by daily infrared satellite imagery during the first 4 days of each event (Figure 4). In J99, the extratropical cyclone and moisture plume remained quasi-stationary in the western Himalaya over 4 days (left-hand panels), illustrating that this was a terrain-locked cyclone in the western notch, as discussed in the introduction. By contrast, due to the more equatorward-displaced trough in M06 (Figure 3b), the cyclone in this event was further south and therefore not as impeded by the mountains. The cyclone progressively moved from west to east over 4 days, appearing to affect different areas of the mountains each day (Figure 4, right-hand panels).

Table 1. Parameterizations Employed in Both Simulations

Microphysics	Thompson [Thompson <i>et al.</i> , 2008],
Surface layer	MM5 Monin–Obukhov [Monin and Obukhov, 1954]
Land surface	Noah [Niu <i>et al.</i> , 2011]
Longwave radiation	Rapid Radiative Transfer Model (RRTM, Mlawer <i>et al.</i> [1997])
Shortwave radiation	Dudhia [Dudhia, 1989]
Boundary layer	Yonsei University [Hong <i>et al.</i> , 2006].
Cumulus	Kain–Fritsch (Kain 2004, outer domain only)

Of particular interest in M06, at 00 UTC 9 March, the extratropical cyclone approached a northeast-southwest oriented convective band over central India that formed in association with enhanced moisture flux convergence between the Arabian Sea and Bay of Bengal (cf. Figure 3d). Twenty-four hours later, the satellite image indicates the extratropical cyclone to have moved east but the convective band to have remained quasi-stationary. Another 24 h later, there is no distinction between them, indicating that they had merged. Therefore, the two systems merged sometime on 10 March (more specifically, between 00 UTC and 06 UTC, as diagnosed by the intervening satellite images, not shown). Thus, from this time onward, the extratropical cyclone intensified the transport of moist and convectively unstable air toward the mountains, contributing to the observed enhancement of precipitation and cloudiness over central-eastern India and the Himalayas. Hence, M06 illustrates an environment in which the extratropical cyclone interacts with tropical moist air (as is typical for events affecting the central Himalaya), whereas the J99 event describes a more typical winter baroclinic-instability event (as is typical for events affecting the western Himalaya). The focus of this paper is on the evolution of the events over 8 days, but detailed analysis of WRF simulations on 6 January 1999 and 10 March 2006 will be presented in section 7.

3. Methodology: Model Runs

Eight-day simulations were performed for 3–11 January 1999 (J99) and 9–17 March 2006 (M06) with version 3.6 of the Advanced Research Weather Research and Forecasting (ARW–WRF, hereafter WRF) model [Skamarock *et al.*, 2008]. Simulations were performed with two domains with one-way interaction (i.e., the nested domain did not affect the outer domain) of 20 km and 6.7 km (3:1 ratio), mapped with a Mercator projection (Figures 1a and 1b). A 4 km grid spacing was tested for part of M06 without any significant difference in the terrain representation or snowfall distribution. Figure 1b shows all the drainage basins that drain across the Himalayan range, where the boundaries are plotted from a gridded file at 6.7 km grid spacing. This representation of the drainage basins illustrates that 6.7 km is sufficiently small grid spacing to capture all the topographical ridges by which the Himalayan drainage basins are defined. Contrast this to the outer domain, which does not capture these topographical ridges (Figure 1a). Hence, this grid spacing is sufficient to determine which drainage basin snow falling in a given location will affect. Because the motivation for this study is to predict moisture availability in given areas for given storms, we therefore consider that 6.7 km is suitable.

The time steps were 90 and 30 s. Initial and lateral boundary conditions were provided by CFSR. The parameterization schemes employed in these simulations are given in Table 1. Maussion *et al.* [2011] found that there is no optimal set of parameterization schemes for WRF simulations over the High-Asia Mountains and that the choice of schemes depends on the specific purposes of the simulation. However, in the current study, equivalent simulations were also performed with different microphysics and boundary-layer schemes (same grid spacing, and initial and boundary conditions) with no qualitative or particular quantitative difference in the resulting snowfall in the mountains.

WRF uses a terrain-following vertical coordinate (with 40 levels used in these simulations from the surface up to 50 hPa), but all output was interpolated onto pressure levels. Eighty model vertical levels were also tested for part of M06 and found to make a negligible difference to the snowfall distribution.

Following Maussion *et al.* [2011, 2014] and the HAR, the model was re-initialized for every day of integration in order to keep the synoptic-scale forcing as close to observations as possible and so that each day of simulation could be compared fairly. Also following Maussion *et al.* [2011, 2014], each day of simulation used for the study consisted of the final 24 h of 36 h simulations, with each day's simulation being initialized at 12 UTC (the day prior to that analyzed) and the first 12 h of integration discarded as model spin-up time.

For each 24 h period of simulation, the model output was taken hourly at 01 UTC, 02 UTC, ... , 00 UTC and the daily mean of all dynamical variables (potential temperature, θ ; geopotential height, z ; wind components, u , v , w ; and all microphysical species represented by the microphysical scheme) thus calculated. For surface-precipitation variables, the accumulation from 00 UTC to 00 UTC was taken. These mean and accumulated values of variables were also calculated for each simulation as a whole.

4. Observational Data Used for Comparison to Model Output

This study is concerned with the spatial and temporal distribution of precipitation when extratropical cyclones interact with the Himalayas. Unfortunately, observations of precipitation at high elevations in the Himalayas are poorly distributed in space and time. Therefore, this study utilizes the satellite-based TRMM gridded-precipitation data set.

TRMM 3B42 [Huffman *et al.*, 2007] is a $0.25^\circ \times 0.25^\circ$ lat–lon grid spanning the entire longitudinal range of the Earth between 50°N and 50°S . The product is obtained by combining infrared and microwave data from multiple satellites to estimate precipitation reaching the surface every 3 h. Station data are used for bias correction. In all figures in this paper in which TRMM data are shown, the WRF data (precipitation and elevation) have been bilinearly interpolated onto the coarser TRMM grid for a fair comparison with this data set.

Because TRMM is itself an estimate of precipitation distribution, we do not present it as a validation of WRF, particularly because TRMM is unreliable in predicting snowfall [e.g., Lang and Barros 2004; Barros *et al.*, 2006; Anders *et al.*, 2006], which is of primary interest to this study. However, as will be shown, for the two events simulated, TRMM agrees with much of the precipitation that WRF simulates at high elevations. While we acknowledge TRMM's limitations, as discussed in section 1 it is the most reliable data set of gridded-precipitation estimates in the Himalayas that is independent of model simulations. We suggest that similarities between WRF and TRMM are due to a common likeness of the true precipitation distribution. In one instance in this paper, satellite data of cloud and snow cover from the Moderate Resolution Imaging Spectroradiometer (MODIS, <http://modis.gsfc.nasa.gov/>) are able to indicate the true distribution of snow that fell over several days.

The APHRODITE [Yasutomi *et al.*, 2011; Yatagai *et al.*, 2012] data set, generated by interpolating available surface station data, was also examined in this study. Because this data set is based on low-altitude stations and there were very few surface observations in the respective regions of interest during each event contributing to the data set, APHRODITE was not considered to be a reliable data set for comparison to the model in this study.

5. Evolution of Snowfall Throughout Simulations

Before examining the dynamics of the simulations, an overview is first provided of the spatial and temporal distribution of snowfall in each simulation. The spatial distribution of accumulated precipitation over each simulation will then be showcased in the next section.

The evolution of snowfall over 8 days in J99 is shown in Figure 5. Each day, the greatest accumulated snowfall occurs where the 700 hPa winds converge against the mountains, depending on where the cyclone is located. The snowfall distribution over the first 4 days changes very little, with one primary snowband exhibited along the southernmost (windward) ridge of the western Himalaya, and another snowband further into the mountains of varying length and intensity. The simulation captures the terrain-locking of the cyclonic winds in the notch (Figures 5a–5d) between the western Himalaya and Hindu Kush Mountains (Figure 1a), as validated by satellite in the left-hand panels of Figure 4. In the last 4 days of the simulation (Figures 5e–5h), the cyclonic circulation and snowfall (and rainfall, not shown) weaken as the cyclone finally moves eastward along the Himalayas.

In M06, by contrast, there is considerable snowfall throughout 8 days of simulation (Figure 6). Unlike J99, this snowfall moves from west to east driven by two successive cyclones that propagate along the Himalayas. The first cyclone can be identified on 9 March by the counterclockwise rotation of the 700 hPa winds approximately between 69°E and 78°E (Figure 6a). This system moves eastward, intensifying southerly moisture transport and snowfall along the central Himalaya (Figures 6b–6e). The location of the cyclone each day agrees with the observed satellite imagery (cf. right-hand panels of Figure 4). A second system, observed

Figure 5. Evolution of daily accumulated snowfall (colors, mm) and daily mean 700 hPa wind vectors (m s^{-1}) in the inner domain of the January 1999 simulation. (a–h) 3–10 January, respectively. Also plotted is elevation (greys, km). Box drawn in Figure 5d indicates location of 6-hourly model data plotted on that day in left-hand panels of Figure 14.

west of the domain on 14 March (Figure 6f), moves along the mountains contributing to additional snowfall in the western and central Himalayas (Figures 6f–6h). Simulation of an event in February 2000 by *Lang and Barros* [2004, Figure 10] indicated a similar evolution of the cyclone and associated precipitation along the Himalayan ridge. However, their simulation only ran for 24 h, which our simulation shows is not nearly long enough for a cyclone to move along the entire ridge. The convex shape of the central Himalaya inhibits the terrain-locking of these systems. On 14 March, the winds associated with both cyclones are exhibited, the rearmost in the western Himalaya and the foremost in the notch in the eastern Himalaya (Figure 1a), where they generate snow for 4 days from 11 to 14 March. Thus, the eastern notch may also trap winds and hence sustain orographic precipitation, as in the western notch in J99.

Figure 6. As Figure 5, but for the March 2006 simulation. Box drawn in Figure 6b indicates location of 6-hourly model data plotted on that day in left-hand panels of Figure 16.

The quick succession of these two cyclones over a few days illustrates how a particularly active jet perturbed sufficiently far south may greatly enhance snowfall in the Himalayas by virtue of more consecutive days of cyclonic activity and hence convergence and snowfall in the mountains. However, unlike in J99, the duration of snowfall in affected areas is generally less than 2 days, due to the transient nature of the system.

6. Spatial Distribution of Accumulated Precipitation Over Simulations

This section exhibits the distribution of accumulated precipitation in the two simulations and compares to TRMM. First, the lat–lon distribution of accumulated precipitation is shown. Then, cross sections across

Figure 7. Total accumulated precipitation over 8 days in January 1999 (colors, mm), comparing WRF rain, WRF snow, and TRMM. (a, b) Also plotted are elevation from WRF (greys, km) and simulation-mean 500 hPa winds. Bold lines annotated “cs1”, “cs2”, “cs3”, and “cs4” show the locations of the cross-section plots in Figures 8 and 12.

accumulated-precipitation maxima illustrate the locations of these maxima relative to mountain peaks and valleys.

6.1. January 1999 Event

In J99, the accumulated rain (Figure 7a) and snow (Figure 7b) from WRF are generally confined to the western Himalaya, reflecting the terrain-locking of the cyclone in this region (Figure 4, left-hand panels, and Figure 5). The location of this precipitation is very similar to the decadal-mean winter-accumulated precipitation in the HAR [Maussion *et al.*, 2014, Figure 2a] and TRMM also reflects this confinement of precipitation to the western Himalaya (Figure 7c). TRMM also agrees with WRF that a relatively small amount of precipitation fell in the eastern notch. Overlaid in the WRF panels are mean 500 hPa winds throughout the simulation, which are very similar to the decadal-mean 500 hPa January winds in the HAR [Maussion *et al.*, 2014, Figure 9]. Specifically, there are cross-barrier winds across the western Himalaya and near-zonally oriented winds south of the Himalayan ridge. Therefore, this flow pattern and precipitation distribution are common for the time of year, not just among extreme-precipitation days.

Four cross sections oriented in the cross-barrier direction across the accumulated precipitation all exhibit a gradual change from WRF rain to WRF snow, moving up the mountains, at about 2 km elevation (Figure 8), demonstrating that this is the melting layer, as will be shown explicitly in section 7. Along all the cross sections, this change from rain to snow is accompanied by a sharper increase in accumulated precipitation (all snowfall rates and accumulations shown in this paper are of snow water equivalent) than that of the rainfall accumulation leading up to the melting layer. This feature is particularly exhibited in cs2 (maximum of 255 mm) and cs3 (maximum of 115 mm), where the mountains are much steeper than in cs1 and cs4. To this effect, Anders *et al.* [2006] showed that a significant positive correlation exists in the Himalayas between snowfall amount and the slope of the terrain in the wind direction.

The WRF and TRMM cross-section curves are generally well aligned (Figure 8) and appear to capture most of the same maxima. In particular, both capture a double maxima of > 100 mm accumulated precipitation (snowfall, according to WRF, which it almost certainly was, given the elevation and time of year), crossed by cs2, arising from the double mountain peak either side of the Kashmir Valley (as is also reflected in [Maussion *et al.* 2014, Figure 2a] indicating that this is a common precipitation feature). Both WRF and TRMM show a third precipitation maximum along cs2, associated with the Karakoram, centered at about (78°E, 35°N)

Figure 8. One-dimensional cross sections at locations shown in Figure 7 showing elevation and comparing total accumulated precipitation over 8 days in January 1999 between WRF rain, WRF snow, and TRMM.

(Figure 7b), although TRMM suggests that it did not extend so far southeast. TRMM reflects the general pattern simulated by WRF that the greatest snowfall is at the first major mountain ridge, with subsequent peaks further into the mountains receiving progressively less snowfall. However, at cs3, which contains the most uniformly rising terrain of the four, there is a single marked snowfall maximum. *Bookhagen and Burbank* [2010] showed that this is a common aspect of precipitation in the Himalayas, namely, that the uniformity of topography determines the number of precipitation maxima in the cross-barrier direction (see their Figure 7).

6.2. March 2006 Event

In M06, the mean location of the moisture plume shown in the right-hand panels of Figure 4 is captured in the WRF rain (Figure 9a) and TRMM (Figure 9c) accumulated precipitation. The near-continuous band of accumulated snowfall (Figure 9b) illustrates the effect of a disturbance sufficiently far south to remain unblocked by the Himalaya, while entraining tropical moisture, producing a large band of convergence, and driving heavy orographic precipitation where cyclonic southerly winds flow normal to the mountains. Over the course of the simulation, the two cyclones traverse the Himalayan ridge, depositing precipitation across the entire windward ridge (Figure 6). This snowfall distribution bears very little resemblance to the decadal-mean accumulated winter precipitation in the HAR [Maussion *et al.* 2014, Figure 2a]. Furthermore, the simulation-mean 500 hPa winds do not resemble the decadal-mean 500 hPa March winds in the HAR [Maussion *et al.* 2014, Figure 9]. The HAR March winds are in fact similar to the HAR January winds: cross-barrier in the western Himalaya and along-barrier in the central Himalaya. However, there is a greater northerly component to the HAR March than January winds, indicating a greater ease for systems to pass along the ridge in March. Therefore, unlike J99, this was an unusual event for the time of year, but cyclones such as in M06 are more likely to pass along the ridge in March than in January.

As in J99, TRMM captures some of the snow further into the mountains, simulated by WRF (cf. Figures 9b and 9c), and TRMM is generally well-aligned with WRF rain and snow in cross-barrier cross sections (Figure 10). Satellite data of cloud and snow cover from MODIS show that, after the passage of the cloud associated with the storm in M06, there were two distinctive bands of snow-cover difference between the

Figure 9. As Figure 7, but for March 2006. Cross section 3 is at the same location as in Figure 7, but other cross sections do not appear in Figure 7, hence are numbered differently.

Figure 10. As Figure 8, but for March 2006 and at cross-section locations shown in Figure 9.

Figure 11. MODIS cloud- and snow-cover data for M06, showing cloud and snow cover on 8 March (before event, top left) and 19 March (after event, bottom left), and pixels at which there was snow cover on 19 March but not on 8 March (right).

beginning and end of the event (Figure 11). These data verify that snow indeed fell in a second band further into the central Himalaya.

WRF generally shows a snowfall maximum between 4 and 5 km elevation (Figure 10). This is slightly higher than in J99 (generally between 3 and 4 km elevation, Figure 8), perhaps because of the greater forcing of precipitation in M06 than J99, as will be illustrated in the next section. *Putkonen* [2004] showed from multi-year station data that winter snowfall in the central Himalaya gradually increases with elevation from 3 km (the lowest elevation at which snow accumulates) at least up to 4.4 km (the highest elevation measured). Therefore, if M06 is representative of the vertical distribution of winter snowfall in the central Himalaya, WRF appears to be reasonably accurate in this respect.

As in J99, WRF simulates large snowfall accumulations atop of steep mountainsides, particularly in cs5 (Figure 10b, 203 mm), where the relief in the central Himalaya is steeper than any of the mountains encountered in J99. This cross section is well within the mean location of the moisture plume (Figures 9a and 9c), suggesting that, in addition to a particularly steep mountainside, snowfall may have been enhanced in this location due to particularly large cross-barrier moisture flux, as is investigated in section 7. WRF simulates a snowfall accumulation of 310 mm along cs7, which is not steep (Figure 10b). The particularly large accumulation in this location appears to be due to persistent snowfall throughout the simulation (Figure 6), rather than particularly high hourly snowfall rates, as is also investigated in section 7.

7. Interaction of Synoptic-Scale Forcing With Mountains

It may be expected, following the ingredients-based approach [e.g., *Johns and Doswell III* 1992], that the snowfall observed and simulated for these two events required some combination of lift, moisture, and instability. Lift is a greater factor for a dynamically driven storm, while instability is more important for a convectively driven (e.g., monsoon) event. Because the precipitation in both J99 and M06 resulted from

Figure 12. Timeseries through January 1999 simulation, showing maximum values every hour along the given cross sections (locations shown in Figure 7) of snowfall rate (black, mm h^{-1}), 700 hPa moisture flux (blue, $\text{kg m}^{-2} \text{s}^{-1}$), and CAPE (red, J kg^{-1}). The r values given above each panel are the Pearson product-moment correlation coefficients of snowfall rate with 700 hPa moisture flux (blue) and CAPE (red).

extratropical cyclones, we may expect that both were ultimately dynamically driven. Figures 5 and 6 suggest this to be true, where the snowfall each day of both simulations is co-located with the strongest cross-barrier 700 hPa winds, depending on the cyclone location. However, the reanalyses and satellite images in Figures 2–4 suggest that due to the warmer waters from above which air was transported toward the mountains in M06 than J99, convective instability may have contributed to precipitation, as indicated by cold cloud tops ahead of the cyclone. This section investigates the evolution of snowfall in each simulation with respect to the possible factors (dynamic versus convective). This analysis is performed by taking cross sections in the cross-barrier direction, as in Figures 8 and 10, but showing dynamical variables. At each cross section, wind components are calculated relative to the cross-section orientation, so that a large x component of wind vectors in cross-section plots indicates strong cross-section-parallel flow at a given pressure level.

7.1. January 1999 Event

Figure 12 shows hourly timeseries throughout J99 at the same four cross sections whose locations are shown in Figure 7 and whose accumulated WRF rainfall and snowfall are shown in Figure 8. At all cross sections, snowfall (black time series) is fairly persistent, reflecting the terrain-locking of the cyclone in this simulation (Figure 5). The maximum hourly snowfall along all cross sections is generally very well correlated through time with the maximum cross-section-parallel (i.e., cross-barrier) moisture flux at 700 hPa (blue time series, column-integrated cross-section-parallel moisture flux was also plotted with very similar results). Therefore, snowfall in the mountains in J99 appears to be heavily dependent on and follow almost inevitably from sufficient moisture and cross-barrier winds, as hypothesized by *Lang and Barros [2004]*.

Furthermore, maximum moisture flux and maximum snowfall rate are generally proportional through time. The scales of these two diagnostics in the time series relative to one another have been purposely set to illustrate that roughly 1 mm h^{-1} snowfall results per $0.01 \text{ kg m}^{-2}\text{s}^{-1}$ cross-barrier moisture flux.

There is also a reasonable correlation in time between snowfall and convectively available potential energy (CAPE, red timeseries), although, unlike moisture flux, there is a great variability in the ratio of CAPE to snowfall rate. In particular, there is generally much greater CAPE along cs3 and cs4 than along cs1 and cs2 (because cs3 and cs4 are further south and experience the flow of high-CAPE air from the Arabian Sea, not shown), but the magnitude of snowfall is similar between all cross-section locations. Furthermore, the winds transporting the moisture are necessary to transport this high-CAPE air from the Arabian Sea to the mountains. Therefore, although the relatively large CAPE near the mountains at times of snowfall must have enhanced vertical motion and hence snowfall rate to some extent, the moisture transport was the crucial factor in generating snowfall and the snowfall rate was mostly determined by the magnitude of moisture flux.

In all the cross sections, particularly cs2 and cs4, 72–96 h (6 January, the date from which J99 was identified using TRMM data—section 2) stands out as receiving the greatest snowfall. Figure 13a shows an observed sounding at Patiala (location marked by “P” in Figure 14a) at 00 UTC, i.e., capturing the flow toward the mountains shortly before this extreme snowfall occurred. A dry layer with southeasterlies up to 700 hPa lay beneath a deep moist layer from 700 to 300 hPa carried by southerlies (i.e., perpendicular to the mountains). Note the single (presumably erroneous) temperature measurement of -81.1°C at 742 hPa. WRF also captures the moist layer and southerlies above 700 hPa at the same time and location but only has the moist layer up to 500 hPa (Figure 13b). Therefore, the snowfall was forced by moist southerlies from 700 hPa upward, as is animated in Figure 14.

Figure 14 shows the evolution of snowfall on this day in the area shown by the box in Figure 5d, with cross sections shown at the location of cs2. This figure shows the passing of the cyclone from the perspective of a particular location in the mountains. At 03 UTC, the location of the cold front is clearly marked just south of the cross section by an abrupt wind shift at 700 hPa, with some low-CAPE air just ahead (i.e., in the cyclone’s warm sector, Figure 14a—the domain maximum of CAPE at this time is 379 J kg^{-1}). There are a number of small-scale snowfall maxima where the southerly winds converge most strongly against the mountains. In addition, widespread light snow is falling over most of the mountainous area. The cross section passes through the most intense snowfall maximum, illustrating that this is where widespread ascent and snowfall between 750 and 350 hPa associated with the cold front is being moved toward the mountainside by the midtropospheric winds (Figure 14b). Unsurprisingly, the base of this snowfall is co-located with the freezing layer at about 750 hPa and at this time most of this precipitation reaches the surface at low elevation as rain (not shown).

The mesoscale response as this frontal ascent encounters the mountainside is readily evident, with a dipole of vertical motion at each major mountain peak and valley, capped by the stable layer between 300 and 200 hPa. The maximum vertical velocity at this cross section throughout the simulation is 3.6 m s^{-1} , indicating that this vertical motion is stable and not unphysical. This is the same order of magnitude as in other simulations of extratropical cyclones interacting with major mountain ranges (e.g., *Barros et al.* [2006], also in the Himalayas; *Rasmussen et al.* [2011], in the Rocky Mountains; *Viale et al.* [2013], in the Andes), and Figure 14b is qualitatively very similar to Figure 19 of *Barros et al.* [2006] and Figure 12 of *Viale et al.* [2013]. Snow falls between the first valley (the Kashmir Valley) and the second mountain peak, leading to the double maximum that was exhibited in the accumulated WRF and TRMM precipitation data (Figure 8b; same cross section). Six hours later, there is widespread snowfall along the cross section, with the intensity and spatial coverage of the snowfall decreasing further downwind, as the vertical motion weakens and the specific humidity decreases (Figure 14d). This feature of the simulation explains the accumulated-precipitation cross sections (Figure 8), where subsequent peaks beyond the first major orographic barrier received progressively less snowfall.

As the cold front moves slightly east (Figures 14e and 14g), the snowfall becomes less widespread along the cross section (Figures 14f and 14h), leaving only the maximum at the first ridge. By this stage, the wind field has smoothed to form a relatively uniform snowband along the ridge (observe transition from Figures 14a–14g). However, the horizontal plots show that a snowfall maximum has remained at this location throughout the day, and there are a number of other locations where this is also evident. These

00 UTC 6 Jan 1999

(a) Observed

(b) WRF

14 Mar 2006

(c) Observed, 00 UTC

(b) WRF, 02 UTC

Figure 13. Skew-T diagrams of observed (left-hand panels) and WRF (right-hand panels) soundings at Patiala (location shown by “P” in Figure 14a) at indicated times and dates. Wind measurements were not available for the observed sounding in Figure 13c.

locations are found where outlying mountains are more exposed to the wind field (in Figure 8, notice how much further upwind the first peak of cs2 lies, compared to the other cross sections). A contrast to this feature of the simulation will be shown in M06.

7.2. March 2006 Event

In M06, snowfall at cs3, cs5, and cs6 (Figures 15a–15c) is less persistent than at the cross sections in J99, reflecting the gradual passages of the two successive cyclones along the mountains, so that no one location receives snowfall for an extended period. The relationship between cross-barrier moisture flux and snowfall rate along these three cross sections is similar to that in J99, in that the two curves are generally very well aligned and are roughly the same relative proportions as in J99 (1 mm h^{-1} snowfall per $0.01 \text{ kg m}^{-2} \text{ s}^{-1}$). The

Figure 15. As Figure 12, but for March 2006 simulation, with cross-section locations indicated in Figure 9.

major difference to J99 is that the moisture flux is generally much larger. This is particularly evident between 24 h and 48 h (during 10 March, the date from which M06 was identified using TRMM data—section 2) in cs5, in which a very abrupt increase in moisture flux occurs. We recall that this was the day of the merger of the cyclone with the convective cloud over eastern India (Figure 4). We also recall that cs5 is within the mean location of the moisture plume (Figures 9a and 9c) and particularly on 10 March (Figure 6b). Furthermore, cs5 crosses particularly steep terrain in the central Himalaya (Figure 10b). The response in terms of snowfall is evident, as the snowfall rate reaches almost 10 mm h⁻¹ (Figure 15b).

An observed sounding is not available near the central Himalaya around this time but one is shown at Patiala just south of the western Himalaya at 00 UTC 14 March (Figure 13c—the day that extreme snowfall occurred in the western Himalaya in M06, Figure 6f) to illustrate the contrast between the air masses of J99 and M06. A saturated atmosphere from the surface up to 500 hPa of 2177 J kg⁻¹ CAPE indicates the far greater moisture content and convective instability of the air advected toward the mountains in M06 than in J99 (cf. Figures 13a and 13c). WRF does not capture this extremely deep moist layer at the same time and location (not shown) but does so just 2 h later (Figure 13d), albeit underlain by a dry layer up to 700 hPa that the observed sounding does not show. Unlike the observed sounding, WRF does not show any CAPE at this location, but Figure 16 (left-hand panels) illustrates the large horizontal variability of CAPE on 10 March, so that areas of zero CAPE are just tens of kilometers from areas of CAPE greater than 1000 J kg⁻¹, as is also the case at the time of these soundings on 14 March (not shown).

Figure 16 shows the evolution of the flow and snowfall along cs5 on 10 March, as this extreme moisture plume flows normal to the mountains, where the location of the left-hand panels is shown by the box in Figure 6b. The greater warmth (indicated by a higher freezing level) and moisture content of the atmosphere upwind of the mountains than in J99 is readily evident (cf. right-hand panels of Figures 14 and 16). As reflected by the observed soundings, there is much greater CAPE ahead of the cold front than in J99 (cf. left-hand panels of Figures 14 and 16—a domain maximum of 2565 J kg⁻¹ at 03 UTC, Figure 16a) and, unlike

Figure 16. As Figure 14, but for March 2006 simulation on 10 March, with area shown in left-hand panels indicated in Figure 6b and right-hand panels showing cs5.

in J99, there is great potential instability at the foot of the mountains, as diagnosed by the cross-section equivalent-potential-temperature (θ_e) contours (cf. Figures 14b and 16b). At 03 UTC, the cold front and strong southerly winds are just to the south of the central Himalaya (Figure 16a). The frontal snowfall is seen in the cross section upwind of the mountainside (Figure 16b) and is much more widespread than in J99 (cf. Figure 14b). The impact of the moisture plume reaching the mountains 6 h later (Figures 16c and 16d) is much greater than in J99 (Figure 14). The snowfall is much more widespread in the along-barrier (Figure 16c) and cross-barrier (Figure 16d) directions. By contrast, in J99, the heaviest snowfall was restricted to the most

outlying peaks. The more stably stratified θ_e contours than 6 h before (cf. Figures 16b and 16d) indicate that potential instability has been released and contributed to the vertical motion.

Over the next 12 h, the cold front and the high-CAPE air ahead of it move east along the Himalayas, moving the snowband east (observe transition from Figures 16c to 16g), so that snowfall at the cross-section location gradually dissipates (observe transition from Figures 16d to 16h). This evolution reflects the transient nature of the storm at individual locations, due to the convex central Himalaya, as opposed to J99 in which the concave western Himalaya allows the snowband to remain quasi-stationary throughout the day (Figure 14).

Cross section 7 is subject to roughly the same cross-barrier moisture flux about 24 h later, but the snowfall rate is noticeably lower than at cs5 (Figure 15d). Throughout the simulation, the ratio of snowfall rate to moisture flux is generally much lower than in the other cross sections. As illustrated in Figure 10, this cross section crosses much less steep terrain than the others, and indeed other cross sections across less steep terrain in this simulation also show lower snowfall rates, proportional to moisture flux (not shown), indicating that steeper relief causes a greater snowfall rate for a given magnitude of cross-barrier moisture flux. However, cs7 also stands out in that the snowfall is much more persistent than in the other cross sections, reflecting the trapping of winds in the eastern notch of the Himalayas (Figure 6), despite the fact that that cyclones of this event were not stationary. Therefore, despite lower snowfall rates than in the central Himalaya, this location is where the greatest snowfall accumulation occurs (Figure 10).

Unlike J99, CAPE magnitude and snowfall rate are generally proportional through time and between different cross-section locations. Specifically, 1 mm h^{-1} snowfall results per 100 J kg^{-1} CAPE. However, over the four cross-section locations, only about half the snowfall maxima in time are coincident with CAPE maxima. Also, there are a few times when there was large CAPE along a cross section but no snowfall, e.g., between 72 and 96 h along cs3. Therefore, although CAPE was far greater in M06 than in J99 and played a greater role in vertical motion and hence snowfall, cross-barrier moisture flux was the dominant factor in M06 as in J99. Thus, we may expect that cross-barrier moisture flux is generally the crucial ingredient for extratropical cyclones affecting the Himalayas, although CAPE can play an enhanced role depending on the state of the background atmosphere.

8. Summary and Conclusions

The scarceness and uneven distribution of observational data pose serious limitations to investigating dynamical and physical processes associated with extratropical cyclones' interaction with the Himalayas. Thus, mesoscale model simulations are required to investigate the structure and evolution of precipitation (typically snow) during these events. Individual events, such as those studied in this paper, may produce a large majority of the total winter precipitation over large areas in the mountains. Therefore, the mesoscale structure of precipitation features associated with any given storm may significantly affect the winter and spring water supply in populated areas downstream.

The subtropical storm track has climatologically shifted northward over southern Asia in recent decades, so that fewer storms pass south of the Himalayas and more become terrain-locked in the western Himalayan notch. Partly due to this trend, some glaciers in the western Himalaya and Karakoram have advanced, while those in the central Himalaya are rapidly retreating. Should this trend continue in the future, there will be even fewer storms affecting the central Himalaya, so that individual storms, when they do occur, are of particular importance for regional hydrology.

Two events in January 1999 and March 2006 were investigated, in which the typical synoptic-scale flow patterns associated with extreme winter precipitation in the western and central Himalaya, respectively, were amplified. These events occurred during moderate and weak La Niña episodes, respectively, and according to TRMM, the winters during which the events occurred were drier than average in the region. However, both events generated the majority of their respective total winter precipitation accumulations over large areas and 1 day during each event was one of the wettest in the western and central Himalaya, respectively, in winters between 1998 and 2013. Therefore, although La Niña episodes may reduce total winter precipitation in the Himalayas, individual intense storms can develop and produce some of the largest precipitation accumulations in the mountains.

In WRF simulations of both events, a band of snow, of intensity on the order of 50 mm day^{-1} (snow-water-equivalent), fell on the southernmost (windward) ridge of the mountains wherever there were strong cross-barrier midtropospheric winds, depending on the location of the cyclone each day. In both simulations, bands of lighter snowfall, but still on the order of 10 mm day^{-1} , formed on ridges further downwind. In the January event, a trough west of the Himalayas meant that the cyclone became terrain-locked in the western Himalayan notch, whereas in the March event, a deep trough further south and east allowed two successive cyclones to move from west to east, with the resulting southerly winds converging against a different stretch of the mountains each day. In the January event, the snowfall was confined to the western Himalaya, while in the March event, there was heavy snowfall along almost the entire ridge. The simulated precipitation compared well with available observational data, TRMM and MODIS, and also other modeling results.

This heavy snowfall occurred at the mountains via the movement of existing widespread frontal snowfall by midtropospheric winds, which, until encountering the mountains, reached the ground at low elevations as rain. However, in the January simulation, this frontal precipitation fell as snow at lower elevations than in the March simulation, due to a lower freezing level. Upon encountering the mountains, the widespread ascent was enhanced and generated a wave-like perturbation in the upper troposphere, leading to descent along downhill stretches and ascent along uphill stretches. As a result, snow fell at multiple peaks further downwind, explaining the bands of accumulated snow further into the mountains.

In the simulations of both events, snowfall maxima (in time) occurred at some location only when there was a maximum (in time) of moisture flux in the midtroposphere in the cross-barrier direction near that location. Therefore, according to the simulations, snowfall in both events was dependent on dynamical forcing. Furthermore, snowfall was generally proportional throughout the simulations to cross-barrier moisture flux. The snowfall in the March event was enhanced for about 2 days following the convergence over India of moisture advected cyclonically from the Arabian Sea with moisture advected anticyclonically from the Bay of Bengal, as is typical for extreme winter precipitation in the central Himalaya, based on reanalysis composites. Following this moisture flux convergence, an abrupt increase in moisture transport against the mountains was evident. Because cross-barrier moisture flux and snowfall rate were generally proportional, there were greater snowfall rates and a more solid snowband during this time than at any other time in either simulation. Furthermore, for part of this time, this strong moisture flux happened to converge against the particularly steep central Himalaya, leading to snowfall rates of up to 10 mm h^{-1} (the same magnitude of moisture flux against the shallower eastern Himalaya led to snowfall rates generally not greater than 6 mm h^{-1}).

There was a greater influence of CAPE in the March than January event, partly because of the time of year and partly because of the warmer waters from above which air was transported toward the mountains, owing to the more southward-displaced trough on the jet, as is typical for events affecting the central than western Himalaya. This influence in the March event was illustrated in a cross-barrier cross section by unstably stratified contours of equivalent potential temperature, which became more stably stratified following snowfall atop of the mountains, indicating the release of potential instability. This result suggests that although cross-barrier moisture flux is the crucial ingredient for winter snowfall in the Himalayas, CAPE is a potentially influential factor according to the state of the background atmosphere.

Both events had features leading to particularly large snowfall accumulations (the stationary cyclone in the January event; the greater cross-barrier moisture flux, greater CAPE, and steeper mountains, plus the fact that there were two successive cyclones, in the March event) and the greatest 8 day WRF snowfall accumulations at individual locations were similar between the two simulations (200–300 mm). However, the heavy snowfall in the March simulation was much more widespread and more total moisture was deposited on the mountains.

These simulations show the relative importance of a number of factors that may affect the intensity and distribution of snowfall when extratropical cyclones interact with high mountains. First, on the synoptic scale, the location of the jet and consequent cyclogenesis, relative to topographic features that have a tendency to terrain-lock a cyclone, may determine whether the cyclone leads to snowfall in one particular region or whether the snowfall is more evenly distributed along the mountains. Second, a particularly active jet may lead to consecutive cyclones, leading to repeated snowfall in the same areas in a short space of time. Third, the interaction of a cyclone with a greater heat and moisture source than it otherwise would have

may lead to enhanced CAPE upwind of and moisture transport against the mountains and hence enhanced snowfall. Finally, depending on where in the mountains the greatest moisture transport happens to be, particularly steep terrain may lead to especially high snowfall rates.

In addition to an improved understanding of individual events, these findings indicate the possible consequences of various aspects of climate change in the region [e.g., Scherler *et al.*, 2011; Dimri and Dash 2012; Ridley *et al.*, 2013]. Future changes in the intensity and location of the jet, changes in upper level disturbances on the jet (whether ENSO-related or otherwise), and changes in sea-surface temperature in areas from which cyclones may transport moisture toward the Himalayas will all be important in determining the distribution of precipitation during such events. An improved understanding of future climate in the region, together with an improved understanding of the precipitation distribution resulting from extratropical cyclones in the Himalayas, will therefore indicate the likely availability of moisture in the region in winters to come.

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