Regional climate model application at subgrid scale on Indian winter monsoon over the western Himalayas

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ABSTRACT: The western Himalayas (WH) is characterized by heterogeneous land surface characteristics and topography. During winter (December, January, and February – DJF), eastward moving low-pressure synoptic weather systems, called Western Disturbances (WDs) in Indian parlance, cause the majority of the precipitation mostly in the form of snow. The interplay between land surface/topography and WDs greatly controls precipitation distribution over the region. This study seeks to evaluate this using a mosaic-type parameterization of subgrid-scale topography and land use (sub-BATS) for regional climate simulation with a regional climate model (RegCM3). The model coarse grid cell size in the control simulation is 60 km while the subgrid cell size is 10 km. This study compares two 22 year simulations (1980–2001) during winter (DJF). The first simulation is without (CONT) and the second is with (SUB) the fine scale subgrid scheme. Representing the fine scale processes using the subgrid scheme SUB experiment simulates reduced precipitation by approximately 2 mm d\textsuperscript{−1} with comparison to CONT experiment. This estimation of reduced and closer to the corresponding observed precipitation is important for regional water budget over the WH which is primarily governed by topographic and land surface disaggregation. Validation with corresponding observations over similar elevations shows that SUB displays an improvement over CONT experiment. This relevant decrease of precipitation in SUB using disaggregation-reaggregation methodology for initial model input fields in subBATS scheme is due to better representation of the WH topography. In case of temperature, SUB experiment displays colder bias (∼2–4°C) than the CONT over the Himalayas. This preliminary finding is important for studying regional water balance, snow melt accumulation in following summer period. Copyright © 2012 Royal Meteorological Society

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1. Introduction

The western Himalayas (WH) is characterized with land surface/topographic heterogeneity. Eastward moving low-pressure synoptic weather systems, called Western Disturbances (WDs) (Pisharoty and Desai, 1956; Rao and Srinivasan, 1969; Singh, 1979; Kalsi, 1980; Kalsi and Haldar, 1992; Lang and Barros, 2004; Dimri and Mohanty, 2009) yield high amounts of precipitation, mostly in the form of snow (hereafter precipitation only refers to liquid precipitation – rainfall), over WH during winter (December, January, and February – DJF). This precipitation is mainly controlled by the interplay of topography and WDs. A number of studies have shown interaction of topography/orography with weather systems and corresponding precipitation over various mountain ranges (Barros and Lettenmaier, 1994; Buzzi et al., 1996; Buzzi and Deser, 1998; Pielke, 2001; Giorgi et al., 2003; Feddema et al., 2005; Bookhagen and Streeker, 2008; Im et al., 2010; Medina et al., 2010). Studies have also shown a relationship between the Himalayan topography and the Indian Summer Monsoon (ISM) (Bhaskaran et al., 1996; Barros et al., 2000; Barros and Lang, 2003; Kripiani et al., 2003; Bookhagen et al., 2005; Anders et al., 2006; Barros et al., 2006; Bookhagen and Burbank 2006). In particular, Anders et al. (2006), for example, has shown a variation of precipitation patterns over a spatial scale of approximately 10 km strongly controlled by the Himalayan topography using TRMM data. Seko and Takahashi (1991) showed the effect of wintertime precipitation on the mass balance of a glacier in the Nepal Himalayas. Many studies have used regional to local-scale models to define the orographic role and variation in precipitation as well (Barros and Lettenmaier, 1994; Buzzi et al., 1998; Das et al., 2003; Barros et al., 2006; Dimri, 2004, 2009; Dimri and Ganju, 2007; Takahashi et al., 2009, 2010).

Despite these different studies on the role of Himalayan topography in defining precipitation, the studies primarily focus on the ISM, and only a few have investigated the relationship between Himalayan land surface/topography and winter weather and precipitation patterns (Barros et al., 2006; Dimri, 2004, 2009; Dimri and Ganju, 2007). Lang and Barros (2004) had provided 30 years...
observation-based climatology over Nepal Himalayas. In this study it is planned to investigate with the modelling efforts if similar kind of climatologies over WH with the regional model is reproduced or not. To begin with, in this paper, we first focus on if there exists any such relationship between Himalayan topography and winter weather and precipitation using regional climate simulations. Also, the role of orographic in precipitation mechanism is defined as an interplay of WDs with topography in terms of precipitation briefly with case study of a WD.

2. Model and experimental design

2.1. Regional climate model

An updated regional climate model (RegCM3) is used (Pal et al., 2007) for this study. The physical parameterization of the comprehensive radiative transfer package of the National Center for Atmospheric Research (NCAR) Community Climate Model, version 3 (Kiehl et al., 1996), the non-local boundary scheme of Holtslag et al. (1999), the mass-flux cumulus cloud scheme of Grell (1993), the resolvable-scale cloud and precipitation scheme of Pal et al. (2000), and the land surface processes of Dickinson et al. (1993) are used in the present model simulation configuration.

For such studies, ideally very high resolution, explicit convection resolving simulations will be preferred (Medina et al., 2010), but computational limitations for multidisciplinary regional climate assessment mandate alternate methodologies involving multiple nesting (Christensen et al., 1998; Leung and Qian, 2003) or parameterization of subgrid-scale processes (Giorgi and Mearns, 1991; Giorgi and Avisser, 1997) are recommended to account for subgrid-scale land surface/topography. In this study, the subgrid scheme of Seth et al. (1994) within a framework of a regional climate model (Pal et al., 2007) is used.

2.2. Subgrid land surface parameterization

Parameterization of land surface heterogeneity here refers to both topography and land use, and can generally be divided into discrete, mosaic methods, or continuous, probability density function (pdf) methods. In the mosaic methods, the model grids are divided into a number of subgrid cells. Land surface calculations are carried out separately for each subgrid and the land–atmosphere exchanges are then reaggregated at the original coarse-grid cell scale. Subgrid cells are based on land use type class (Avissar and Pielke, 1989; Koster and Suarez, 1992), topographic elevation class (Leung and Ghan, 1995), or local land use or topography (Seth et al., 1994). Giorgi and Avisser (1997) discussed the advantages and limitations of these different ways of representing land heterogeneity. In pdf methods, the heterogeneous variables are represented via analytical or empirical pdfs and relevant processes are integrated over the appropriate pdf (Entekhabi and Eagleson, 1989; Avissar, 1991, 1992; Dumenil and Todini, 1992; Famiglietti and Wood, 1994a, 1994b; Sivapalan and Woods, 1995; Giorgi, 1997a, 1997b). Regardless of methods used, earlier studies have shown that the subgrid-scale heterogeneity in topography and land use conditions can profoundly affect climate and the surface energy and water budgets, especially at the regional and local scales.

In this study we have implemented an augmented version of the mosaic-type scheme within the RegCM3 (Giorgi et al., 2003). Each model grid cell is subdivided into a regular subgrid of N cells with equal area. Each cell has its own specific topographical elevations, vegetation class, and soil type for which independent land surface calculations are carried out. For these calculations, solar and infrared downward radiative fluxes, precipitation, near-surface air temperature, water vapour, wind speed, pressure, and density are taken as input from atmospheric models. Once, the land surface processes are computed, the Biosphere Atmosphere Transfer Scheme (BATS) returns the albedo, surface upward infrared, momentum, sensible heat, and latent heat flux to the atmospheric model. Therefore, for atmospheric model simulation, which is run on the coarse grid, and BATS on the fine subgrid, the atmospheric input to BATS are disaggregated from the coarse grid to the subgrid and after calculations, BATS information are re-aggregated back at the atmospheric model grid scale.

This disaggregation for the temperature field is completed according to the subgrid elevation difference and expressed as

\[ T_{i,j}^{sg} = \bar{T} + \Gamma_T (\bar{h} - h_{i,j}^{sg}) \]

where \( sg \) is subgrid and the overbar refers to the coarse grid, \( T \) and \( h \) denote the surface air temperature and topographical elevation, respectively, and \( \Gamma_T \) is the mean atmospheric lapse rate.

An important constraint for the disaggregation is

\[ \bar{h} = \frac{1}{N} \sum_{i,j} h_{i,j}^{sg} \]

that is, the coarse grid elevation is equal to the mean of the subgrid elevations. This suggests that the model grid point surface air temperature is equal to the average subgrid near-surface atmospheric temperature. Similarly, surface pressure and density are computed based on standard gas law using subgrid-scale temperature, whereas subgrid-scale near-surface air specific humidity is computed assuming a constant near-surface relative humidity across subgrid points equal to the coarse grid scale relative humidity.

2.3. Experimental design

The western Himalayan land surface/topography modulates wintertime weather and profoundly controls precipitation distribution. Therefore, it is imperative to understand such interplay between land surface/
topography and WDs in defining climatic variability in precipitation and temperature. Knowledge of these two fields is of utmost importance for regional development, such as water, agriculture, and tourism. To assess such interplay we conducted and analysed two sets of model runs: a control run (CONT), in which the fine scale BATS scheme is not used and therefore the land surface has the same resolution as the atmosphere, and a fine scale subgrid-scale based run (SUB), in which the BATS scheme is used. The domain covers an area from the Gulf of Aden to north India using a Lambert Conformal projection centred over central Asia, with grid cells of 60 km × 60 km size for CONT (coarse grid cell) and 10 km × 10 km for SUB (Figure 1(a) and (b)).
Table I. Model configurations used in the present study.

<table>
<thead>
<tr>
<th>Model configuration</th>
<th>Details</th>
</tr>
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<tbody>
<tr>
<td>No. of horizontal grid points</td>
<td>iy = 51 (along y-direction)</td>
</tr>
<tr>
<td>No. of vertical levels</td>
<td>Kz = 23</td>
</tr>
<tr>
<td>Distance between grid points</td>
<td>ds = 60 km</td>
</tr>
<tr>
<td>Centre of latitude</td>
<td>31 N</td>
</tr>
<tr>
<td>Centre of longitude</td>
<td>70 E</td>
</tr>
<tr>
<td>Horizontal grid scheme</td>
<td>Arakawa-Lamb B grid staggering</td>
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<tr>
<td>Time integration scheme</td>
<td>Split explicit</td>
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<tr>
<td>Surface parameters</td>
<td>BATS1E</td>
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<tr>
<td>Map-projection</td>
<td>Lambert conformal</td>
</tr>
<tr>
<td>Sea surface temperature</td>
<td>OISST</td>
</tr>
<tr>
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<td>NNRP2</td>
</tr>
<tr>
<td>Model physics</td>
<td>Grell</td>
</tr>
<tr>
<td>Cumulus scheme</td>
<td>Frisch and chappell</td>
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<tr>
<td>Lateral boundary condition</td>
<td>Relaxation (exponential)</td>
</tr>
<tr>
<td>Planetary boundary layer scheme</td>
<td>Holtslag</td>
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<tr>
<td>Large scale precipitation scheme</td>
<td>SUBEX</td>
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<tr>
<td>Ocean flux parameterization scheme</td>
<td>Zeng</td>
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<tr>
<td>Pressure gradient scheme</td>
<td>Normal way</td>
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<tr>
<td>Lake model</td>
<td>No</td>
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<tr>
<td>Tracer/chemistry model</td>
<td>No</td>
</tr>
</tbody>
</table>

Each coarse grid cell is divided into 36 subgrid cells. Topographical information to obtain the two grids is taken from a 2 min resolution global dataset produced by the United States Geological Survey (USGS). Fine scale fractional land use cover information over different surface types are taken from the Global Land Cover Characterization (GLCC) dataset. Atmospheric fields from the National Centre for Environmental Prediction (NCEP) Reanalysis II (Kanamitsu et al., 2002) are used as initial and lateral boundary conditions and the National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation Sea Surface Temperature (OISST) dataset is used for SST over the ocean areas. NCEP reanalysis is at 2.5° horizontal resolution with 6 hourly time interval. The OISST dataset is at a 1° horizontal resolution with a weekly time interval. In experimental strategies, CONT and SUB simulations are made continuously for 22 years (1981–2002) with initial 4 months (1 September 1980 to 31 December 1980). Simulations are not included in the analysis in the first 3 months (1 September 1980 to 30 November 1980) in order to allow model spinup. A brief model configuration is presented in Table I.

2.4. Observational dataset

In this paper, discussions are mainly focused to assess the relationship between land surface and topographical feedback and significant climatic variables, that is, winter precipitation (again to mention that precipitation is referred for liquid precipitation-rainfall only) and temperature. Three observational reanalysis datasets used for the model evaluation and performance are (1) 0.5° resolution global land datasets developed by the Climate Research Unit (CRU) of the University of East Anglia (New et al., 2000), (2) 0.25° resolution gridded precipitation dataset for Asia (APHRODITE) (Yatagai et al., 2009), and (3) 2.5° resolution NCEP Reanalysis II (Kanamitsu et al., 2002). Also, for more in-depth analysis in-situ observations at three stations, viz., Gulmarg (lat 34°30′00″N, lon 74°29′00″E, alt 2800 m), Bhag (lat 32°16′33″N, lon 77°09′03″E, alt 2192 m), and Base Camp (lat 35°11′49″N, lon 77°12′28″E, alt 3570 m) of the Snow and Avalanche Study Establishment (SASE), Chandigarh, India is used. These stations represent different climatic and geographic conditions of the WH and have the longest records available. This study uses two different observationally based gridded precipitation data and hence it is important to mention that the number of stations per grid cell is available for APHRODITE. This information is very useful to determine to what extent the gridded precipitation is determined from station data or derived using interpolation between the stations. Also, climate over WH is colder and drier than other Himalayan regions therefore daily time resolution of APHRODITE becomes more reliable than monthly time resolution of CRU. Both the datasets are over land but with different space resolution: APHRODITE has finer space resolution than CRU. In APHRODITE, daily observation records are collected from meteorological and hydrological organizations of countries from Asian monsoon regions. It is also important to mention that though APHRODITE datasets include very few stations inputs while preparing its preparation, but uses topographic information (vertical and horizontal both). This is one of the important issues over the Himalayan region. Methodologies to construct precipitation reanalysis are described in Yatagai et al. 2009 and New et al. 2000 for APHRODITE and CRU, respectively. Apart from the above, various fields, including outgoing long-wave radiation cloud cover fraction, wind, surface pressure, vertical velocity, and so on are analysed to assess the impact of land surface/topography on precipitation and temperature.

3. Results and discussion

The role of the WH land surface/topography with precipitation and temperature is discussed in the following paragraphs. In addition, interannual variability is also discussed. Comparison with observed data sets and associated error analysis, in terms of root mean square error (RMSE) and correlation coefficient (CC), is carried out to assess the model’s skill.

3.1. Topography

The WH is characterized by heterogeneous land surface/topography with high peaks and low valleys running along the NW–SE direction. WH consists of five
major subranges in a cascading manner—the Pir Panjal, Great Himalayas, Zanskar, Ladakh, and Karakoram. This topographic distribution plays a crucial role in defining the spatial distribution of precipitation and is discussed in subsequent paragraphs. Figure 1(a)–(c) shows topographic distribution from the model simulation (CONT and SUB experiments) and observed topography (GTOPO30_10 min) which is used in the model initialization, respectively. Figure 1(b) for the SUB experiment, pronounced topographical information as compared to the CONT experiment (Figure 1(a)) and has topography generally similar to the 10 min observed shown in Figure 1(c). These additional features correspond to more pronounced peaks and valleys represented in the SUB experiment will provide finer-scale details of climatic variables over the WH region. By having subgrid information, resolvable fine-scale features are expected to be better explained due to parameterization of coarse grid surface heterogeneity at the fine-scale surface subgrid. Also, spatial distributions of climatic variables over the windward and/or leeward side of various mountainous ranges in the WH will possibly be simulated better. The model sensitivity with topography is mainly discussed over a more complex region of the WH (Figure 1(d)). For detailed investigations, one in-situ observation from each region, A (Gulmarg), B (Bhang), and C (Base Camp), as shown in Figure 1(d) and referred in Section 2.4 is considered. Figure 1(e)–(g) and Figure 1(h)–(j), correspond to the latitudinal and longitudinal cross-sectional topography across the WH at 30°N, 35°N, 40°N, 70°E, 75°E, and 80°E, respectively. These cross sections represent strong topographic gradients across the E–W and N–S directions.

3.2. Surface pressure

First, simulated and observed (analysis) mean surface pressure climatology over the WH is analysed to assess the model sensitivity. Figure 2(a)–(c) represents seasonal (DJF) mean surface pressure climatology corresponding to the CONT and SUB experiment and the NCEP verification reanalysis, respectively. The model simulates similar surface pressure distribution over the central Indian region as seen in the corresponding verification analysis. A similar distribution of central surface pressure of the order of approximately 500 hPa over the central Himalayan region is also seen. This illustrates that the model could also simulate mean surface pressure well over high and low surface regions. In addition, low surface pressure maxima of the order of approximately 700 hPa at and around 34°N 68°E in verification analysis could also be simulated by the model. Apart from this, pressure gradients are produced across the variable WH topography. This comparison with corresponding verification analysis demonstrates that the model has the ability to reproduce surface pressure features with accuracy. Further, the SUB experiment (Figure 2(b)) produces more additional surface features than the CONT experiment (Figure 2(a)). Better surface pressure field is simulated in SUB experiment than CONT experiment as fine grid scale disaggregation procedure is used. Whether this translates into better resolution of circulation, mesoscale convection and precipitation are assessed in the following sections.

3.3. Circulation features

The zonal wind profile during boreal winter (DJF) at 200 hPa illustrates significant dynamical large-scale features. It exhibits stratified structure of the mean subtropical westerly jet (SWJ) centred along 28°N (Figure 3(a)) as seen in the corresponding verification reanalysis (Figure 3(b)) with an approximate 3 m s⁻¹ lower magnitude. The stratified structure depicting meridional wind distribution at 200 hPa, Figure 3(c) shows centres of maximum northerly (∼30°N) and southerly (∼25°N) winds with higher intensity than the corresponding verification analysis (Figure 3(d)). Apart from this there is slight shift in these zones. Further, the vertical structure of the model-simulated mean zonal wind at 30°N cross-sectional latitude (Figure 3(e)) and corresponding verification analysis (Figure 3(f)) shows a similar vertical structure in the upper atmosphere with a centre core extending at 200 hPa. However, the lower atmospheric model field shows higher vertical shear particularly at the surface, around 80°E, where easterlies are produced. In
case of vertical distribution of mean meridional components, (Figure 3(g,h)), similar vertical structures is reproduced but with slight westward shift of zone of maxima. The structure of strong northerly maxima at 300 hPa and around 65° E and southerly maxima at 200 hPa and around 82° E could be simulated though with slightly higher meridional wind values. A strong discontinuity in the wind surface along 72° E could also be well simulated. Here as well, strong shear are produced in the vertical distribution of mean meridional wind than seen...
in the corresponding verification of the meridional wind field. Also, at the surface around 70°E and 80°E, strong zones of northerly and southerly wind fields are seen. Sawyer (1956) has suggested vertical variations in vertical velocity such that upward motion immediately above the hill is reversed and replaced by downward motion upwards and again by upward motion at some higher level. The role of Himalayan topography is apparent in these circulation patterns. The influence of topography can be attributed primarily to localized disturbances of the vertical structure of the atmosphere. Such disturbances exert important control, acting either as barriers, elevated heat sources, and sinks or as concentrated areas of high roughness (Smith, 1979; Barros and Lattenmaier, 1994).

To analyse the wind fields in more detail, model and corresponding verification reanalysis winds are studied at different levels. At 200 hPa, the model-simulated (figures not presented) and NCEP wind field with streamline show similar flow patterns with some stronger gradients produced by the model (CONT) over the central Himalayan region (25°N 83°E). Similar flow patterns are simulated in CONT at 500 hPa as well. At 850 hPa, somewhat stronger boundary layer flow patterns are produced by the model. The CONT run could not reproduce the exact flow direction and strength. However, the southwesterly flow at 38°N 62°E, northeasterly to northwesterly flow over the head of the Arabian Sea, southwesterly flow over the Indo-China region, and northeasterly flow over the Indo-Gangetic Plain are simulated with higher wind speeds. Also, there is a cyclonic circulation at 850 hPa centred at 30°N 75°E which is also observed in verification reanalysis with less amplitude. To assess it further, mean seasonal divergence fields are evaluated. The CONT seasonal (DJF) divergence field at 200, 500, and 850 hPa is presented in Figure 4(a), (c), and (e). The corresponding divergence field based on NCEP verification reanalysis is shown in Figure 4(b), (d), and (f). Broad-scale divergence and convergence regions at various atmospheric levels are simulated by the model. Upper tropospheric convergence and lower tropospheric divergence is simulated over the Indian continent. Interestingly, at 850 hPa strong divergence over the Himalayas and convergence over southwest China is produced. It is to be noted here that verification analysis is at coarse resolution and hence model experiments, being at finer resolution, represent the role of land surface and topographic distribution better. Similar dynamical behaviour is seen in the mean seasonal (DJF) omega field (Figure 5). As seen in Figure 5(a) and (c) and (b) and (d) the mean seasonal omega field for the CONT experiment and corresponding NCEP verification reanalysis at 200 and 500, respectively. In the case of omega at 200 and 500 hPa, the large-scale omega field is simulated over the India subcontinent. In comparison, the impact of the Himalayan topography is clearly visible. Localized cells of positive and negative omega are seen in the model which may correspond to high elevation peaks and low valley points across the Himalayas. Particularly, at lower levels upward motion along the Himalaya is seen, otherwise the model does simulate the omega distribution as seen in the verification reanalysis field. Here the role of orography on the spatial scale is important as the cloud and precipitation forming process are dependent on such local feedbacks. Such flows can determine the potential of strengthening/weakening of convective formation over narrow regions within the planetary boundary layer with sharp horizontal moisture gradients (McGuire, 1962; Peckham and Wicker, 2000; Medina et al., 2010).

3.4. Precipitation

Winter precipitation is an important variable over the WH as it is the main source of water for most of the north Indian rivers during the ablation period. In this study model output-based liquid precipitation (rain) and not solid precipitation (snow) is compared with APHRODITE fields as later also correspond to rain only. It is important to inform that during winter substantial amount of precipitation takes place in snow form. So, to see this aspect water equivalent observed at station point are compared with model liquid precipitation derived at that point location and discussed too. Figure 6 shows a 22 year seasonal (DJF) mean simulated (Figure 6(a,b), CONT and SUB) and observed (Figure 6(c,f) CRU, and APHRODITE) precipitation along with their differences (Figure 6(d,e,g,h) for CONT-CRU, SUB-CRU, CONT-APHRODITE and SUB-APHRODITE). CRU observations show that the mean seasonal (DJF) precipitation over the WH is generally wet with the centre of maxima lying over the Hindukush region. Model simulations illustrate the role of topographically-induced precipitation patterns along the Himalayan region. The model over-estimates maximum seasonal precipitation by approximately 4 mm d⁻¹ showing a wet bias particularly over high mountainous regions. Both experiments produce two precipitation maxima: one over the Indo-Pak region and another over the Afghanistan region around 37°N 75°E and 38°N 72°E, respectively. Error analysis of two experiments with CRU and APHRODITE (Figure 6(d,e,g,h)) shows that the SUB experiment produces better spatial distribution and less bias than the CONT experiment with correlation coefficients of 0.73 and 0.81 with CRU and 0.77 and 0.89 with APHRODITE in the CONT and SUB experiments (significant at the 95% confidence interval). The spatial correlation coefficients are calculated after the observations are aggregated onto the model grid and including all the points in the interior domain. Corresponding RMSEs are 0.68 and 0.52 with CRU and 0.44 and 0.32 with APHRODITE in the CONT and SUB experiments. This illustrates that finer spatial resolution in the observational gridded data enhances accuracy in model simulations. These error analysis and, comparisons of corresponding model biases show that model has reasonable ability to simulate precipitation distribution over the Himalayan range. Spatial distribution is closer to the APHRODITE than CRU except over high elevation points. This can be attributed to the fact that higher
density of observation points is used in preparation of APHRODITE than CRU. Overall comparison shows two maxima of mean seasonal (DJF) precipitation, as seen in APHRODITE observations, could be reproduced by the model. The model produced a much closer simulation to the APHRODITE observations with less wet bias. To assess model behaviour, model simulations are analysed over WH (Figure 1d) (Figure 7). Error analysis with the CRU observation shows that the SUB experiment produces approximately 2 mm d$^{-1}$ less precipitation than the CONT experiment and the difference between the two experiments shows that the SUB experiment simulated spatial precipitation distribution better than the CONT experiment over high elevation of the WH. Here we can
see that the SUB experiment produced less precipitation over high peaks as compared to the CONT experiment and hence less wet bias by approximately 2 mm d$^{-1}$. Bias with APHRODITE observation shows that except over high elevation peaks model simulates less wet bias over the Himalayas. Here it is not easy to exactly quantify how much and with which observational gridded data model comparison is better. This is to be noticed that in APHRODITE weighting function is imposed not only to calculate horizontal distance, but also local topographical features such as elevation and mountain slopes, which improves the orographic precipitation patterns (Yatagai et al., 2009). To look into exact performance of CONT and SUB experiments over the WH (Figure 1(d)), Figure 7 illustrates detailed spatial precipitation distribution. Figure 7 depicts regional climatological estimates of wet biases in CONT and SUB experiments with respect to CRU and APHRODITE and also relative improvement in SUB. Here we can see that SUB experiment simulated climatology shows better results than CONT experiment produces climatology as subgrid topographic parameterization is used. Figure 7(c) shows that SUB performs better in high topographic regions of the Himalayas during winter than CONT. This provides relative importance of using SUB as it improves representation of topography in model. This finer scale topography discretization calculates initial model input fields at finer model horizontal resolution with vertical topographic gradients and then reaggregates back these fields for final atmospheric model simulations. This gives an extension to the Lang and Barros (2004) where they have discussed storm climatology over central Himalayas. In the present work it is suggested that SUB reproduced closer climatology than CONT over the WH. To analyse this precipitation difference further, latitudinal and longitudinal cross-sectional distribution of mean seasonal (DJF) simulated (CONT and SUB) and observed (CRU and APHRODITE) precipitation is studied and presented (Figure 8). To better interpret the rainfall changes, model-simulated (CONT and SUB) and observed (GTOPO30_10 min) cross-sectional topography is also plotted in the figure. Here, Figure 8 is primarily produce to give an idea of spatial variability in precipitation amount across the WH. Interaction between topography and precipitation is clearly depicted and realistic precipitation values are produced by the model as a result of the topographical feedback. In the case of the latitudinal cross section, model results produced winter precipitation till similar elevation as by observed fields.
Beyond this elevation, in both the experiments and observations, no precipitation is seen. Elevation increase along the mountain upslope is characterized by a decrease in temperature and resistance to the upslope flow. These results in cloud and precipitation formation and rise of air will shed the condensed moisture along the upslope (Yasunari, 1976; Singh et al., 1995). However, the model does produce greater precipitation at higher elevation points and less precipitation at low valley points than the observed with model-simulated precipitation peaks as seen in the observations. At a few places, lead/lag in occurrence of maxima of precipitation values are seen in SUB and CONT due to land surface and topography – as in SUB experiment finer topography is used than in the CONT experiment. In the case of the longitudinal cross-sectional mean seasonal (DJF) precipitation, a similar distribution is seen. It is to be noted that the longitudinal cross-sectional topography is more variable than the latitudinal topography over the WH. However, it still needs more investigations, error analysis with precipitation types and weather cells to finally assess the possible elevation limit of precipitation. A similar finding is proposed by Anders et al. (2006) over the Himalayan region using TRMM precipitation records. Over Nepal Himalayas, Barros et al. (2000) and Lang and Barros (2002) has shown annual precipitation above 4000 m in Maryasandi meteorological catchments. In addition, Barros et al. (2004) has proposed different orographic–land–atmospheric interactions based on scaling analysis of cloudiness during ISM over Nepal Himalayas. They determined linkages between the Himalayan topography and the space–time variability of cloudiness as a function of elevation. Bookhagen et al. (2005) showed that violent rainstorms conquered orographic barriers and
Figure 7. 1980–2001 Averaged DJF precipitation (mm d\(^{-1}\)) for the (a) CONT minus CRU, (b) SUB minus CRU, (c) SUB minus CONT, (d) CONT minus APHRODITE, and (e) SUB minus APHRODITE over WH (Figure 1(d)) region. This figure is available in colour online at wileyonlinelibrary.com/journal/joc.

Figure 8. 1980–2001 averaged DJF precipitation (mm d\(^{-1}\)) (on left hand axis) and topography (*1e\(^{-2}\) m) (on right hand axis) along a latitudinal cross-section at (a) 30\(^\circ\) N, (b) 40\(^\circ\) N, and (c) 45\(^\circ\) N and a longitudinal cross section at (d) 70\(^\circ\) E, (e) 75\(^\circ\) E, and (f) 80\(^\circ\) E. CONT-Pr and CONT-Topo correspond to control precipitation and topography, respectively. Similarly SUB-Pr and SUB-Topo correspond to subgrid experiment. CRU-Pr corresponds to CRU precipitation. APH-Pr corresponds to APHRODITE precipitation. This figure is available in colour online at wileyonlinelibrary.com/journal/joc.
configuration snow drift accumulation and rain shadow effects are not treated. These are important mountainous physical processes which need explicit driving mechanism in the model physics (Leung and Ghan, 1995). Fractal interpolation based orographic rainfall disaggregation scheme has shown about 50% improvement of total precipitation amount in quantitative precipitation forecasting (Bindlish and Barros, 2000). Figure 9(a) shows that for low intensities the simulated distributions match the observed station data reasonable well. But in the middle and high intensity range, the model overestimates. Distribution of model precipitation overestimation is clearly seen in the intensity distribution. While comparing this distribution with seasonal distribution of precipitation at station level it is seen that the model’s higher value events are not always well matched with that in the observations. Hence, it can be stated that both experiments simulate the lower value events with higher accuracy than the higher value events. Additionally, the SUB experiment simulates higher values with more accuracy than the CONT experiment. Especially, during winter precipitation generation mechanisms are mainly of dynamical nature over WH. It is evident from the comparison between observation and model result. Small differences across the simulation are essentially due to the internal model variability (Giorgi and Bi, 2000). In contrast, during summer a greater effect of the subgrid scheme is expected as disaggregation procedure is used for convective precipitation and because of which greater forcings by the surface fluxes are expected by subgrid scheme. Further, to analyse the domination of either low or extreme rains over the Himalayan domain shown in Figure 1(d), the power spectrum analysis is carried out. The power spectrum of the seasonal (DJF) precipitation distribution from the observation (APHRODITE) and model runs (CONT and SUB) is presented in Figure 9(b). The APHRODITE observation is chosen as the model runs could simulate similar spatial distribution of seasonal precipitation over the WH. Apart from the seasonal mode of approximately 90 d, figure illustrates a dominant mode of 7–14 d at the synoptic scale in winter seasonal precipitation. This mode, as well, is simulated in both the CONT and SUB runs although with larger power. This higher power is due to a wet bias in the model simulation which has been discussed in the previous paragraphs. This synoptic-scale (7–14 d) duration periodicity is linked to occurrence and lifecycle of cyclone storms (WDs) during winter (DJF). A possible reason for the above difference, Figure 9(a), could be that during winter most of the WH received precipitation in the form of snow and measured in cm d at station. This amount after multiplied with corresponding observed density (or with average density) is presented as corresponding snow water equivalent in mm d\(^{-1}\) in the present study. This change in nature of precipitation may bring certain uncertainty while comparing with corresponding observationally gridded precipitation data. Also, there seems to be similar intraseasonal variability in observation and model results as well (it is under study).
3.5. Surface air temperature

During winter (DJF), surface air temperature in the WH region often remains below sub-zero in the WH. Figure 10 presents the spatial distribution of model-simulated (DJF) (CONT and SUB), observed (CRU), and mean seasonal (DJF) temperature and their biases. Observations show the lowest negative temperature of the order of approximately 258 K over the Indo-Tibetan region around 32°N 80°E and the highest positive temperature over central India. Also, two zones of minimum lowest temperature are noted; one over the Hindu Kush region and another over the Tibetan region. Model experiments could simulate the spatial mean seasonal (DJF) temperature distribution over the WH and reproduce zones of lowest minima well. In the case of temperature simulations, the SUB experiment produces more detailed spatial distribution over the WH. The model tends to underestimate temperature over high elevation areas in the WH and hence shows a cold bias except over the region of maximum topographic heterogeneity where the bias is typically less than 2°C. Further, spatial distribution of seasonal mean temperature in the CONT and SUB simulations along with observations over the WH region (Figure 1(d)) is investigated (figures not presented). It is evident that the SUB simulation, although retaining the basic patterns of the CONT simulation (and thus the similar bias patterns) exhibit much finer resolution information. The bias distribution shows that the SUB experiment could produce better simulation than the CONT experiment over high elevation points. The subgrid scheme mostly redistributed the grid-scale temperatures according to the fine and coarse topographical information without introducing systematic differences. Here, the SUB simulation could produce the lowest temperature range that could not be produced well in the CONT simulations. Also, the effect of subgrid
topography could be clearly seen in the spatial temperature distribution across the WH. Though overall, topographically induced spatial temperature distribution is seen, but this discrepancy can be partially due to the relatively low density of high elevation stations in the observed dataset, where the data probably underestimates temperature over the WH. In addition, Figure 11 illustrates temperature distribution along the latitudinal and longitudinal cross section in these experiments with observation. Comparison with topographic distribution shows that temperature decrease with an increase in elevation or vice versa is well simulated across latitudinal and longitudinal cross sections and the model could reproduce mean seasonal temperature variation with topography over the WH. The SUB experiment provides a finer temperature distribution when a high density of high and low elevation points are included in the subgrid scheme. To analyse the interannual variability of mean seasonal (DJF) temperature, a time series anomaly for the CONT and SUB experiment with observation (CRU and station) is studied (figures not presented). The temporal evolution shows similarities between the simulated and observed fields. Although the model tends to underestimate the winter temperature by 2°C, time series phase coherence is seen. It shows that the model could reproduce interannual variability on the seasonal time scale. Comparisons with station observations show that the model could reproduce the seasonal anomaly at the station level which corroborates the fact that while simulating area averages, many high and low density point values get filtered out and hence model simulations produce biases. Further, when comparing model-simulated and observed (station) temperature pdfs (Figure 12) a shift towards lower values and hence a cold bias was found. Pdfs based on model simulations are slightly wider as well, indicating a higher variance when compared to pdfs based on station observations. While looking for the reasons for this cold bias and shift towards the colder mean value, it is seen that model simulations could not reproduce corresponding mean values. Although the model could well reproduce variances in the pdfs, it could not simulate observed mean values which are higher (Figure 12(a–c)). However, the description of how temperature varies with elevation is crucial. Lundquist and Cayan (2007) and Lundquist and Rochford (2007) in their studies over central Sierra Nevada, California region demonstrated a link between westerly wind, temperature, and mountain slopes. They illustrate that strong westerly winds are associated with relatively warmer temperatures on the east slope and cooler temperatures on the west slope of the Sierras and weaker westerlies are associated with the opposite patterns. Mahrt (2006) proposed that most of the temperature gradients along the slopes are bounded by a maximum value where a further increase of temperature gradients is restricted by the redistribution of heat by thermally driven slope circulations. Zangl (2005) has shown that over the Alpine valley system, the primary mechanism for controlling the persistent air pool is the pressure gradient imposed by (geostrophically balance) ambient air flow and hence the
probability for persistent cold air pools mainly depends
on the ambient wind direction. Such a distribution is yet
to be seen over the WH where cascading valleys and
peaks makes understanding of the interaction of topog-
raphy with weather more complex. McCutchan (1983)
studied temperature and humidity interactions with moun-
tain slopes and found on mountain slopes temperature
becomes lower and humidity becomes higher than the
adjacent free air nearby during the day, whereas Tang
and Fang (2006) demonstrated that the lapse rate along
the southern and northern slope vary considerably over
Mt. Taibai.

This study shows that producing precise simulations
of wintertime climate over the WH is difficult. The
complex nature of the topography and its interaction
with the environment makes it challenging. However
consideration of fine scale processes through subgrid-
scale model parameterization helps to develop realistic
simulation fields over the Himalayan region.

Precise determination of the origin of precipitation and
temperature biases is complex as topography and cloudi-
ness are key factors in defining distribution of these
variables. Figure 13 compares the distribution of the frac-
tion of cloud cover (in percentage) from the CONT
experiment (Figure 13(a)) and the CRU observation
(Figure 13(b)). NCEP outgoing longwave radiation is
also presented in Figure 13(c). During DJF, as expected,
CRU observations locate maximum cloud cover over
the Indo-Pak-Afghanistan region and minimum cloud
cover over the west-central Indian region. The CONT
experiment reproduces spatial distribution over the region
properly, but overestimates cloud cover by showing two
centres of maxima oriented along the Himalayas and
underestimates it over the west-central Indian region.
These over- and underestimations could be one of the
reasons for the model bias but more sensitivity experi-
ments over the Himalayan region are needed. Outgoing
longwave radiation also shows higher values over the
west – central Indian region and lower values over the
WH. This distribution suggests that during winter (DJF),
higher cloudiness conditions exist for lesser outgoing
longwave radiation. The effect of such distributions on
precipitation and temperature needs to be investigated for
bias assessment. Barros et al. (2004) has shown catego-
ration of storms on spatial distribution over the Nepal
Himalayas based on satellite imageries. Similar study
based on satellite imageries with corresponding storms
will be carried out during winter over WH.

23 January 1999

In the present section interactions of large scale circu-
lations with WH topography is discussed to explain the
mechanism of orographically forced precipitation during
a WD. This WD (21–23 January 1999) is chosen as it
gave widespread large amount of precipitation over the
A. P. DIMRI AND D. NIYOGI

WH. This WD caused for, rain/snow, cold waves and subsequently avalanche in mountainous regions across the WH. Synoptically, in brief, this WD is observed extending from Afghanistan to Indo-Pak region as a trough at 500 hPa and low pressure at surface on 21 January. On 22 January, this surface low remained stationary with associated upper air circulation up to 700 hPa and trough at 500 hPa extended up to 300 hPa. On 23 January, pressure trough remained at same location, but pressure gradient steepened over the region and associated upper air cyclonic circulation extended up to 200 hPa tilting westward with height. By 24 January this situation decayed comparatively.

Due to the length of the paper, mechanism of orographic precipitation associated with this WD is presented in brief. Figure 14(a)–(c) illustrate 24 h model simulated cumulative precipitation received at 0000 UTC of 21, 22 and 23 January 1999, respectively. Figure 14(d)–(f) illustrate corresponding observed precipitation based on APHRODITE. Spatial distribution shows zone of maximum precipitation oriented all along the Himalayan topography – which is seen in corresponding observation too (here we are not discussing model biases as detail discussion with reasons is presented in the receding sections). Heterogeneous spatial precipitation distribution is seen during 3 d of the WD in simulation and observation both. Similar precipitation structures are seen in simulation and corresponding observation. Reason for such distribution is looked into. Figure 15 presents longitude and vertical cross sectional distribution of meridional wind and air specific humidity at 34°N latitude. This cross section is chosen as highest topographic variability is seen across this latitude. Higher vertical wind shear in lower troposphere and stronger meridional wind from surface to 500 hPa along the Kashmir valley (∼73°E) is discernible during WD. Also, contrast in meridional wind from surface to 200 hPa along approximately 67°E is seen. Along the valley topographic boundaries wind become slower and in the mid of the valley wind is higher. Also, increase in air specific humidity up to mid troposphere over the valley is clearly visible which is lower along the valley topographic boundaries. To assess it further, spatial distribution of wind surface drag stress over the WH is studied and presented in Figure 16. It illustrates drag produced in large scale flow by orography. Distribution shows role of...
orography in modifying large scale surface flow. Dominant orographic forcing is seen along the WH. Higher peaks of drag stress is seen along the orientation of the orography during all 3 d of the WD. Zone of maximum precipitation (Figure 14) is seen lagging behind towards northwest of the zone of maximum drag stress. It suggests that precipitation generally follows once orographic forcing comes into existence over the WH. Particularly, on 22 January two maxima of surface drag stress corrobo-rate for forming two maxima of precipitation behind. To investigate it further, cross-sectional distribution of 10 m zonal and meridional wind and corresponding surface...
drag stress along 34°N latitude is presented in Figure 17. Prominently it show weakening of meridional wind along the valley bottom (∼72°N) and strengthening of it along the valley boundaries. Zonal wind also weakens along the valley bottom but with a distance lag with meridional wind. However meridional wind peaks along the windward side and dips along the leeward side of the valley boundary. Exact reverse is seen in case of zonal wind.
4. Conclusions

The performance of RegCM3, driven with NCEP reanalysis II in reproducing observed climatology, seasonal cycle, and interannual variability of precipitation and temperature is evaluated. Also, the interplay of the regional climate over the western Himalayan region as modulated by the complex land surface/topography is analysed.

Results show that the region has several challenges in simulating the different atmospheric features. However, the regional model was able to capture circulations influencing the WH well, such as the SWJ in the upper troposphere during winter. It is important as during winter most of the synoptic scale systems, WDs, travel in this larger scale westerlies. True representation of SWJ in model simulation will corroborate with true representation of winter storms. In present case study two-way coupling with large scale circulation is not employed and it is mainly controlled by the boundary conditions. The inability to accurately reproduce surface circulations in areas where topography dominates local winds is one of the main systematic biases in regional climate runs. Simulated seasonal precipitation climatology has a wet bias when compared to the observations. It is important to note that consideration of fine scale subgrid-scale processes is important in realistically simulating regional features. As such, the SUB experiment precipitation climatology is closer to the observations than the CONT experiment. The topographic gradients are better represented by the disaggregation procedure used in the SUB experiment. While the model has a propensity for overestimating precipitation over high mountainous regions, the maxima are well represented over the complex topography. A lower range of precipitation with the synoptic-scale dominant mode is well captured by RegCM3. In the simulated seasonal temperature climatology, a cold bias is seen over the Himalayan region. The model simulation has captured the temperature changes with topography fairly well. The SUB experiment displays a larger cold bias (\(\sim 2-4 \, ^\circ\text{C}\) more than in the CONT experiment) in lower altitude regions but its performance improves over the CONT experiment with an increase in altitude. Interplay of large scale circulation with WD and its manifestation on precipitation mechanism shows that how surface drag stress due to orographic forcing controls it.

In summary, the simulations analysed in this study indicate a reasonably good performance of regional climate model RegCM3 over the complex Himalayan topography. Though difficult to clearly distinguish, present set of model configuration, viz., the choice of domain, involved schemes, and parameter values, have cumulatively worked for good performance. It is also noteworthy that present dynamical downscaling experiments could provide improved understanding of interactive aspects of topography-precipitation-temperature over the WH. In case of precipitation, the synoptic-scale mode and lower range of precipitation could be suitably accounted for in the present setup of RegCM3; however, the higher range

Figure 17. Longitude-pressure cross section vertical distribution at 34\(^\circ\)N latitude of model simulated 10 m zonal wind \((\text{m s}^{-1})\) (yellow contour), 10 m meridional wind \((\text{m s}^{-1})\) (red contour) and uv surface drag stress \((\text{m s}^{-1})\) (black contours) at 0000 UTC during active WD. (a) 21 January 1999, (b) 22 January 1999, (c) 23 January 1999. (Left-hand side-vertical axis corresponds to value from winds and stress and right-hand-side vertical axis corresponds to the topography *1e-2* m). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

These distributions of winds produce highest surface drag stress along the rising relief of the orography and lowest surface drag stress along the downslope relief of the orography. Such drag stress due to orographic forcing produce precipitation over the WH region.
of precipitation was not. This inability to assess extreme precipitation thresholds needs to be carefully examined as they can be lost in coarser-scale climate studies involving rainfall changes and water resources in the Himalayan region.

Additional studies are required to investigate the biases in precipitation and temperature fields which may be due to the large CONT domain (60 km) and SUB domain (10 km). Therefore, apart from the schemes and parameters, analysis at a higher model resolution with difference domain sizes is needed. Additional studies pertaining to model based storm climatology, precipitation – elevation mechanism, role of soil type/vegetation and precipitation – evaporation cycle is under way. It is important to mention that though regional climate model shows robustness nature over the WH at mountain scale but still over complex topography it does not provide satisfactory results at event scale. Also, issue pertaining to the precipitation type and quantity is uncertain within the regional climate model framework.

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