Wintertime land surface characteristics in climatic simulations over the western Himalayas

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Wintertime regional climate studies over the western Himalayas with ICTP-RegCM3 simulations through 22 years has shown systematic biases in precipitation and temperature fields. The model simulated precipitation shows systematically wet bias. In surface temperature simulations, positive and negative biases of 2°–4°C occurred. Experiment without (CONT) and with subBATS (SUB) shows that later scheme performs better, especially for precipitation. Apart from the role of topography and model internal variability, land surface characteristics also have profound impact on these climatic variables. Therefore, in the present study, impacts of land surface characteristics are investigated through cool/wet and warm/dry winter climate by CONT and SUB simulations to assess systematic biases. Since SUB experiment uses detailed land-use classification, systematic positive biases in temperature over higher elevation peaks are markedly reduced. The change has shown reduced excessive precipitation as well. Most of the surface characteristics show that major interplay between topography and western disturbances (WDs) takes place along the foothills rather than over the higher peaks of the western Himalayas.

1. Introduction

The sinks and reservoirs of heat and moisture are influenced by the diverse land surface characteristics of the western Himalayas. This land surface heterogeneity and varying topography can profoundly affect the land surface–atmosphere momentum, water, and energy exchanges. Evaluating such influences in land–atmosphere system is an important component of any regional and global climate modelling system (Xue *et al* 1996; Bonan 1997; Pan *et al* 1999; Zeng *et al* 1999). Also, varying surface vegetation strongly influences the intensity and time-space structure of the Asian Summer Monsoon (Sato *et al* 1989); and changing soil moisture conditions affect rainfall in the monsoon region (Milly and Dunne 1994; Dirmeyer

and Shukla 1996; Kanae *et al* 2006). A number of studies have demonstrated that simulations of surface climate are very much dependent on the formulation of their land surface schemes like soil moisture (Mitchell and Warrilow 1987; Meehl and Washington 1988; Gallimore and Kutzbach 1989), snow cover (Washington and Meehl 1986; Yamazaki 1989; Barnett *et al* 1989), and desertification (Xue and Shukla 1993). Versegny (1991) and Versegny *et al* (1993) have experimented with soil and vegetation model in coupled runs and suggested inclusion of vegetation stomatal resistance for improved model results. Giorgi (1997a, 1997b) has described theoretical and experimental framework representation of land surface processes in climate models. Jacob *et al* (2000) have shown that

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continental climate responds more than global climate with changing land surface characteristics. Ohta *et al* (2001) have illustrated impact of land-use variation on energy and water budget in eastern Siberia. Robock *et al* (2003) have examined the relationship between interannual variation of strength of Indian summer monsoon and land surface conditions (snow cover, soil moisture, surface air temperature and atmospheric circulation) over Eurasia. A similar study was conducted by Xue *et al* (2004) for East Asia and West Africa monsoon development. Gupta *et al* (2005) have studied the role of deforestation on Indian summer monsoon and revealed that there is a differential increase in precipitation rate in northern and decrease in southern India. Another study by Yasunari *et al* (2006) have shown systematic increase of precipitation over the Asian monsoon region based on land surface effect and Tibetan Plateau effect. Pielke *et al* (2007) have provided an overview of regional land-use and land cover change and its impact on rainfall. Over the complex Tibetan topographical region, Xu *et al* (2009) have examined the role of evaporation and water cycle in a lake basin.

In a way land-use and land cover changes (LUCC) affect dynamics and thermodynamics of the atmosphere through the control of energy budget. Representation of the LUCC in most of the land cover datasets are prepared based on various maps, atlases, field observations, etc. (Mathews 1983; Oleson *et al* 1997; Copeland *et al* 1996). Due to different resolutions and observation techniques, these datasets have inherent problems related to accuracy, spatial resolution and updating (Townshend *et al* 1991). Remote sensing techniques have helped to generate land surface details over the Himalayas having a diverse climate (wet–arid–desert) (Sellers *et al* 1995; Smith *et al* 1997; Jin *et al* 2002). Xue *et al* (1996) revealed reduced systematic bias illustrating profound impact of land properties in land–atmosphere interaction through numerical experiments. Pan *et al* (1999) found responses of surface temperature, moisture and wind to LUCC are nearly independent of climate regimes, but those of precipitation to LUCC are highly dependent on climate regimes. Similarly, various studies on the impact of LUCC on regional climate response have been carried out all over the world (Copeland *et al* 1996; Bonan 1997; Lean and Rowantree 1997; Heck *et al* 2001; Pielke *et al* 2007).

It is very difficult to quantify regional climate change over the western Himalayas as suitable land cover datasets are not available and land–atmosphere interactions are highly nonlinear. Also, coupling of land surface parameterization scheme with various dynamical models needs realistic land surface data which includes land cover map, vegetation phenology, soil/snow properties, soil/snow

moisture, etc. These are the serious issues over the western Himalayas than any other region. Further, the western Himalayas is characterized with land surface/topographic heterogeneity and during winter (December, January and February – DJF) 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 Halder 1992; Lang and Barros 2004; Dimri and Mohanty 2009) yield high amounts of precipitation, mostly in the form of snow, over the western Himalayas. This precipitation is mainly controlled by the interplay of topography and WDs.

In the present study, two extreme winter climates, cool/wet and warm/dry, are chosen from 22 years (1981–2002) simulation to assess systematic biases in CONT and SUB experiments (Dimri and Niyogi 2011). Despite a number of similar studies conducted worldwide (Xue *et al* 1996; Fu 2003; Pielke *et al* 2003), no investigation with modelling efforts is attempted for winter season (December, January and February – DJF) over the western Himalayan region. In section 2, a brief description of data and model experiment is given. Section 3 describes the results and discussion of CONT and SUB sensitivity experiments with main focus on land surface characteristics and their impact on surface variables. Finally salient features of the study are concluded under section 4.

2. Model and experimental design

2.1 Regional climate model

An updated regional climate model (RegCM3) is used (Pal *et al* 2007) for the present 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 nonlocal 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. In addition, in the present study, the subgrid scheme of Seth *et al* (1994) within a framework of a regional climate model (Pal *et al* 2007) is used. This kind of parameterization has improved effects chiefly over mountainous regions (Giorgi *et al* 2003; Dimri and Ganju 2007; Dimri 2009; Im *et al* 2010).

2.2 Subgrid land surface parameterization

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:

$$T_{i,j}^{sg} = \bar{T} + \Gamma_T (\bar{h} - h_{i,j}^{sg}) \quad (1)$$

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 Γ_T is the mean atmospheric lapse rate.

An important constraint for the disaggregation is

$$\bar{h} = \frac{1}{N} \sum_{i,j} h_{i,j}^{sg} \quad (2)$$

i.e., 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.

BATS was originally developed to provide realistic land-atmosphere interaction for GCMs. It divides earth's surface and soil into 18 and 12 types respectively to represent realistic properties of land surface. The model has a vegetation layer, a snow layer and surface soil (differentiated into upper layer (0.1 m), the root layer (1.0–2.0 m) and deep layer (3.0 m)). Vegetation is represented by the fraction of vegetation, leaf area index, and stem area index and vary seasonally as a function of the soil temperature. The canopy and canopy foliage temperature are calculated diagnostically

via an energy balance formulation including sensible, radiative, and latent heat fluxes. Snow depth is prognostically calculated from snowfall, snowmelt and sublimation. Snow is the assumed form of precipitation, if the temperature of the lowest model level is below 271 K. Soil types are defined as a function of land cover because of inaccessibility of proper soil database. Land-atmosphere interaction is calculated with a bulk aerodynamic formulism and soil temperature is calculated by a generalized force-restore method (Deardoff 1978). Sensible heat, water vapour, and momentum fluxes at the surface are calculated using a standard drag coefficient formulation based on surface-layer similarity theory. The drag coefficient depends on a surface roughness length on the atmospheric stability in the surface layer.

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 analyzed 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 centered over central Asia, with grid cells of 60×60 km size for CONT (coarse grid cell) and 10×10 km for SUB. 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-hour 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 1980–31 December 1980). First 3 months (1 September 1980–30 November 1980) simulations are not included in the analysis to allow model spin-up.

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, i.e., winter precipitation and temperature. Two 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. Methodologies to construct precipitation reanalysis for CRU is described in New *et al* 2000. (2) 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, long. 74°29'00"E, alt. 2800 m), Bhang (lat. 32°16'33"N, long. 77°09'03"E, alt. 2192 m), and Base Camp (lat. 35°11'49"N, long. 77°12'28"E, alt. 3570 m) of the Snow and Avalanche Study Establishment (SASE), Chandigarh, India is used.

3. Results and discussion

Figure 1 shows distribution of land-use over full model domain of study. A detailed land-use distribution is provided by SUB experiment (figure 1b). Land surface processes are controlled by surface

roughness and albedo. Different land surfaces will have different roughness length and albedo. Table 1 illustrates vegetation types and their corresponding roughness length and albedo. The roughness length and albedo are specified for each vegetation type described in BATS. Roughness length directly affects fluxes of momentum, heat and water vapour between boundary layer and the free atmosphere. Therefore, different vegetation types will impact heat budget. In addition, a vegetated surface generally has a lower albedo and snow surface has a higher albedo. BATS represents the albedo of grid box as an average of soil albedo, vegetation albedo, and snow albedo weighed by the respective area function. The albedo of snow-covered lakes and permanent snow/ice is representative of the albedo of pure snow. It is presumed as SUB experiment uses BATS, it will have explicit information about roughness and albedo than the CONT experiment. Jin *et al* (2002) have discussed albedo variation with land-use and suggested that under snow free conditions, land shortwave albedo rarely exceed 0.2 except for the barren land. Over the Himalayan region, during boreal winter, ground condition is mainly of forests with underlying snow and on an average albedo are lower than 0.30 (Betts and Ball 1997).

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. Figure 2 shows 22 years simulated (figure 2a, CONT and SUB) and observed (figure 2c, CRU) seasonal (DJF) mean precipitation, CRU observations show that the mean seasonal (DJF) precipitation over the WH is generally wet with the centre of maxima lying over the Hindukush

Figure 1. Land-use used in the present study for (a) CONT and (b) SUB experiment.

Table 1. *Vegetation/land cover and corresponding roughness length and albedo.*

BATS vegetation/ land cover	Roughness length (m)	Vegetation albedo (wavelength $<0.7 \mu$)	Vegetation albedo (wavelength $>0.7 \mu$)
Crop/mixed farming	0.08	0.10	0.30
Short grass	0.05	0.10	0.30
Evergreen needleleaf	1.00	0.05	0.23
Deciduous needleleaf	1.00	0.05	0.23
Deciduous broadleaf	0.80	0.08	0.28
Evergreen broadleaf	2.00	0.04	0.20
Tall grass	0.10	0.08	0.30
Desert	0.05	0.20	0.40
Tundra	0.04	0.10	0.30
Irrigated crop	0.06	0.08	0.28
Semi-desert	0.10	0.17	0.34
Ice cap glacier	0.01	0.80	0.60
Bog or marsh	0.03	0.06	0.18
Inland water	0.0004	0.07	0.20
Ocean	0.0004	0.07	0.20
Evergreen shrub	0.10	0.05	0.23
Mixed woodland	0.80	0.06	0.24
Forest/field mosaic	0.3	0.06	0.18

Mean seasonal (DJF) Precipitation (mm/d)

Figure 2. 22 years (1981–2002) 3-month (DJF) average precipitation (mm d^{-1}) for (a) CONT experiment, (b) SUB experiment, and (c) CRU observation.

region. Model simulations illustrate the role of topographically-induced precipitation patterns along the Himalayan region. The model overestimates maximum seasonal precipitation by $\sim 4 \text{ mm d}^{-1}$ 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 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 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 root mean square errors are 0.68 and 0.52 with CRU 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. Error analysis with the CRU observation shows that the SUB experiment produces $\sim 2 \text{ mm d}^{-1}$ 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.

In case of surface air temperature during winter (DJF) over the WH region, it often remains below subzero in the WH. Figure 3 presents the spatial distribution of 22 years seasonal (DJF) mean in model simulation (CONT and SUB: figure 3a, b) and observation (CRU: figure 3c). Observations show the lowest negative temperature of the order of $\sim 258 \text{ K}$ 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 Hindukush 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 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) exhibits 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, 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.

To assess probable reasons for these systematic biases, present study investigates impacts of LUCC on wintertime variability under two extreme seasons chosen from 22 years wintertime model simulations of CONT and SUB experiments. These are cool/wet winter of 1986–1987 and warm/dry winter of 2000–2001 (hereafter referred to as Case-I and Case-II). Model performance for these two seasons has shown higher biases (figure 4), making them interesting cases to study. Surface climate and land cover types under these two selected seasons are analyzed to establish probable reasons for higher systematic biases. Compared to the observed, a dry seasonal precipitation for Case-I and a wet seasonal precipitation for Case-II is simulated by the model, as shown in figure 4. The model has also simulated warm seasonal temperature for Case-I and cool seasonal temperature for Case-II. These two extreme climate regimes are therefore chosen to investigate the systematic biases in model simulations.

Figure 5 illustrates difference in the simulated surface temperature and precipitation from CONT experiment and CRU observations. In simulations,

Figure 3. Same as figure 2, but for temperature ($^\circ\text{C}$).

Figure 4. Normalized anomalies for winter mean CRU precipitation (left hand axis) and temperature (right hand axis) over western Himalayas (30°N ; 72°E to 37°N ; 82°E) with model biases.

Figure 5. Difference fields in winter 3-month (DJF) averaged surface temperature (above: $^{\circ}\text{C}$) and precipitation (below: mm d^{-1}) between CONT simulation and CRU observation for cool/wet winter 1986–1987 (a and c) and the warm/dry winter 2000–2001 (b and d).

cold bias of the order of 2° – 4° C over Himalayan and Tibetan regions and warm bias over south China are observed (figure 5a, b). Though, location and magnitude of winter mean biases in surface temperature are almost independent of season (Pan *et al* 1999), they are lower during cool/wet winter (Case-I) than warm/dry winter (Case-II). It is evident that difference is due to the complexity associated with thermal regimes. Land cover types of positive and negative bias regions are mostly prescribed by desert (Tibetan Plateau) and ice cap/glacier (Himalayas). Core location of warm biases remains consistent with simulation time. There is a possibility that simulated temperature responds slowly to diurnal forcings, leading to a damping of sensible and latent heat fluxes. These factors will become more crucial when soil is relatively wet and subjected to strong heating (Verseghy 1991). Also, daily maximum and minimum surface temperatures over the Himalayan region are significantly changed due to subgrid disaggregation. This decrease, particularly over the Himalayan/Tibetan

region, has led to higher cold bias. This high systematic bias in temperature can be attributed to the prescribed land-use distribution in BATS which is a function of soil properties. Similar findings are proposed by Xue *et al* (1996) over central United States and Suh and Lee (2004) over East Asia.

In case of model overestimation of precipitation over the western Himalayas, realistic simulations are produced over higher elevation points. Impact of LUCC on precipitation is highly dependent on climate regimes but also on time of simulations (Liu *et al* 1994; Dai *et al* 1999). It may have some relation to the complexity of dynamic and thermodynamic processes involved with moisture and precipitation as pointed out by Pan *et al* (1999). A wet bias over the Himalayan region and dry bias over Tibetan Plateau is seen for precipitation (figure 5c, d). Model simulated higher precipitation ($\sim +4$ mm d^{-1}) over western Himalayan region and lower precipitation (~ -1 mm d^{-1}) over Tibetan Plateau. Higher elevation peaks over the Himalayan region has permanent ice (frozen

Figure 6. Difference fields in winter 3-month (DJF) averaged precipitation (mm d^{-1}) between SUB and CONT simulation (above) and between SUB simulation and CRU observation (below) for cool/wet winter 1986–1987 (a and c) and the warm/dry winter 2000–2001 (b and d).

ice, glaciers, etc.) and during winter WDs (Dimri and Mohanty 2009) yield fresh amount of snow over these regions which is variable in nature. Apart from that, land-use over the Himalayan regions is either irrigated crop or semi-desert or ice cap/glacier, therefore the core of temperature bias is dependent upon land-use and precipitation amount. Figure shows that Case-I received higher seasonal precipitation cooling down the surface resulting in lower systematic bias in temperature

which happens to be comparatively higher during Case-II. Visual comparison shows that the model has less systematic bias for temperature and more bias for precipitation in Case-I than in Case-II. Patterns of precipitation bias are highly related to the movement, intensification, and residence time of the WDs (Lee and Suh 2000). However, higher precipitation over northwest Himalayas and central India is seen in Case-I which is simulated by the model. Precipitation distribution in CONT and

Figure 7. 3-month (DJF) averaged 10 m wind (m/s) with streamlines in SUB experiment (above); CONT experiment (middle) and difference (SUB minus CONT – below) for cool/wet winter 1986–1987 (a, c and e) and the warm/dry winter 2000–2001 (b, d and f).

SUB experiments is shown as difference plots in figure 6. In both winters, SUB experiment simulates higher precipitation over the foothill region of the Himalayas and less precipitation over high elevation peaks of the Himalayas (figure 6a, b). A comparison of the simulated and the observed data shows that SUB experiment has wet bias over western Himalayan region and dry bias over Tibetan Plateau (figure 6c, d). However, the SUB experiment produces detailed distribution of precipitation with less bias as compared to CONT experiment (figure 5c, d). Strong moisture gradients are seen along and across the Himalayan topography. Most of the high precipitation cells are formed along the northwest to southeast Himalayan orientations. The change in terrain slopes produces small fluctuations in the virtual potential temperature. As the terrain slope increases, the virtual potential temperature value cools. All these associate factors produce high relative uncertainty in precipitation over complex Himalayan region (Tian and Peters-Lidard 2010). Kitoh (2002) has shown that with removed lapse rate effect and mountain uplift, continents become warmer resulting in reduced moisture transport with fewer clouds and drier surface whereas monsoon regions became cooler due to presence of moisture and hence increased cloud and wet surface.

The 10 m wind and streamlines corresponding to these two seasonal extremes in two simulations and difference are presented in figure 7. A similar discontinuity wind surface lying along the Himalayas is observed, however, two different distinct circulatory features are seen during these two winters.

In Case-I, a well marked circulatory convergence over the Afghanistan region and in Case-II, an organized circulation over and along the Himalayan foothills is seen. Along the foothills, from the northwest to southeast, 10 m easterly winds dominate in Case-II, which becomes north-northeasterly and moves towards the Arabian Sea. It is difficult to make any quantitative judgement here; however it indicates that surface winds play a dominant role in determining seasonal dynamical character as well. Prominently two phenomena, either blocking or forcing the flow around the mountain, take place under such conditions. Smolarckiewicz and Rotunno (1989, 1990) have shown that in case of Froude number ($Fr \geq 1$), most of the flow will go over the topographic obstruction, and for smaller Fr airflow around the obstruction will be favoured. Blocking the flow is associated with low wind speed and high atmospheric stability which strongly affects atmospheric dynamics in the Himalayan region (Barros and Lettenmaier 1994). In case of WDs travelling in large scale westerlies, this situation will become complex as cold air jets travel faster than lower blocked warm air. This creates circumstances for embedded convection to occur (Smith 1982). Jarraud *et al* (1987) have shown an increase of precipitation in mid-latitude under such conditions. Impact of roughness length change on surface wind is almost independent of climate regimes and the month of simulation (Pan *et al* 1999). Decrease of roughness length over the foothill of the Himalayas results in increase in surface wind speed due to inclusion of subBATS in SUB experiment. Also, increase of roughness

Figure 8. Difference field in winter 3-month (DJF) averaged 850 hPa wind (m/s) with streamlines between CONT experiment and NCEP reanalysis with specific humidity (shaded) in CONT experiment for cool/wet winter 1986–1987 (a) and the warm/dry winter 2000–2001 (b).

length over the high elevation points of Himalayas result in decrease in surface wind speed. In case of surface wind, radiative cooling creates favourable conditions for cold air pool formations. Surface snow has high albedo resulting in reduced heating during the day, and with the possibility of radiative cooling due to clear night provides atmospheric layer with high static stability (Zangl 2003, 2005). Interaction of 10 m wind with topography primarily determines surface circulations. This relationship between wind and moisture is presented in figure 8. Here wind streamline corresponds to

difference in 850 hPa wind between CONT experiment and corresponding NCEP verification analysis. Higher specific humidity is seen over Pak to Indo-Gangetic Plain and central Indian region, and low specific humidity is seen over Tibetan region. In Case-I excess specific humidity is observed over Gujarat region to central Indian region. In case of 10 m wind well marked convergence zone over Indo-Pak region with dominant southwesterly flow over central Indian region is seen. This southwesterly flow suggests higher moisture incursion from Arabian Sea in Case-I. Visual comparison of

Figure 9. Difference fields in winter 3-month (DJF) averaged surface drag stress ($\times 10^{-3}$ m/s) (above); sensible heat flux (W/m^2) (middle) and evapotranspiration (mm/d) (below) between SUB and CONT experiment for cool/wet winter 1986–1987 (a, c and e) and the warm/dry winter 2000–2001 (b, d and f).

Case-I and II shows that 10 m wind patterns in former case are more conducive for higher specific humidity than the latter case. Over higher peaks of the Himalayas, insignificant changes are discernible as realistic precipitation rarely crosses ~ 4000 m elevation (Bookhagen and Burbank 2006).

Higher drag could force moistened air to rise higher along the mountain slope. Therefore, surface drag stress is evaluated to assess role of orographic forcings. Figure 9(a, b) illustrates difference field of SUB minus CONT surface drag stress. Though similar spatial distribution is seen along the Himalayan region, higher surface drag towards the windward side and lower surface drag towards the leeward side of the Himalayas is simulated in SUB experiment than CONT experiment. Comparison shows slightly higher stress in Case-I. Higher surface drag is produced along the foothills than over high elevation peaks. This illustrates that higher land surface–atmosphere interaction occurs along the foothills than over the high elevation peaks. Further, considering the wind speed and heat fluxes, differences in sensible heat flux for both winters are shown in figure 9(c, d). Increase in surface wind provides favourable

condition for exchange of sensible heat flux between the land surface and atmosphere. Figure illustrates that sensible heat flux decreases over foothills of the Himalayas for both winter extremes. Although broad spatial difference pattern look similar, the impact of LUCC is stronger on cool and wet winter (Case-I), as lower sensible heat flux distribution is simulated by SUB experiment. Apart from these land cover over foothills of the Himalayas has changed from deciduous needle leaf (CONT experiment) to mixed wood (SUB experiment), hence stomatal resistance and roughness length is reduced and evapotranspiration is noticeably increased (figure 9e, f). This changed effect is quite discernible in foothills. Higher evapotranspiration along the foothills and lower evapotranspiration over high elevation peak of Himalayan is seen in Case-II than in Case-I. To assess this in detail, seasonal precipitation–evapotranspiration (P–E) distribution during these climatic regimes is evaluated and presented in figure 10(a, b) for CONT and figure 10(c, d) for SUB experiment. Figure illustrates positive values over high elevation and negative values along and over foothills of Himalayas and over central Indian

Figure 10. Difference fields in winter 3-month (DJF) averaged precipitation (mm/d) and evapotranspiration (mm d⁻¹) in CONT (above) and SUB (below) simulation for cool/wet winter 1986–1987 (a and c) and the warm/dry winter 2000–2001 (b and d).

Figure 11. Same as figure 10, but for difference between ground temperature ($^{\circ}\text{C}$) and 2 m air temperature ($^{\circ}\text{C}$).

Figure 12. Time series of surface temperature derived at point location with station observation (at Gulmarg) and area (30°N ; 72°E to 37°N ; 82°E) averaged precipitation with station observation (at Gulmarg) for cool/wet winter 1986–1987 (a and c) and the warm/dry winter 2000–2001 (b and d). The 5-day moving average is applied.

region. Comparison of these two winters shows that Case-I is having lesser but distributed evapotranspiration over foothill and central Indian region, which in Case-II is higher with many minima lying

along the foothill region. Even over high elevation region less P–E is seen in Case-II than in Case-I. Corresponding ground temperature and 2 m temperature differences are presented in figure 11. In

Case-I, higher 2 m temperature over higher elevation region dominates than in Case-II (figure 11a, b). Over foothill regions reverse is seen. Over central India high ground temperature is observed in Case-I. Due to this kind of spatial temperature distributions, spatially distributed evapotranspiration over foothill and central Indian region took place in Case-I which in Case-II remain centered along foothill region. It suggests that role of land surface and atmosphere interactively decides the P–E cycle over the region.

In order to see diurnal variation in surface temperature at Gulmarg (station observation and corresponding lat./long. value from model simulations) and area (30°N, 72°E to 37°N, 82°E) averaged precipitation (APHRODITE observation and corresponding value from model simulations) in these two climate extremes pentad time series is prepared and presented in figure 12(a, b) and figure 12(c, d) respectively. In case of diurnal variation of temperature, in both the winters, patterns are simulated by both the experiments – SUB simulating lower values than CONT. Whereas in case of precipitation, SUB has simulated more realistic precipitation values over the Himalayan regions. Higher precipitation events are well simulated by both the experiments in two winter extremes.

4. Conclusions

Land-use characteristics are studied on regional wintertime climate over the western Himalayas with and without BATS for cool/wet and warm/dry climate. Interactions between land-use on surface variables – temperature and precipitation – are studied to ascertain reasons for systematic biases in climatic simulations. Response of land surface–atmosphere interactions remain independent of seasons, however influence of BATS land-use is seen on temperature and precipitation. With BATS, reduced cooling is seen over Himalayan region with more realistic precipitation simulations. Though the effect of BATS on surface winds is not observed, its role on different seasons is clearly discernable. In summary, the simulations analyzed in this study indicate a reasonably improved understanding of role of land-use with regional climate model RegCM3 over the western Himalayas. It is difficult to clearly state about combination of the choice of domain, involved schemes, and parameter values. However, present dynamical downscaling experiments provided improved understanding of interactive aspects of land-use–topography–precipitation–temperature over the western Himalayas. Additional studies are still 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 different domain sizes is needed.

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