The Dominant Role of Snow/Ice Albedo Feedback Strengthened by Black Carbon in the Enhanced Warming over the Himalayas

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(Manuscript received 23 October 2018, in final form 9 June 2019)

ABSTRACT

An obvious warming trend in winter over the Tibetan Plateau (TP) in the recent decades has been widely discussed, with studies emphasizing the dominant effects of local radiative factors, including those due to black carbon (BC). The Himalayas are one of the largest snowpack- and ice-covered regions in the TP, and an ideal area to investigate local radiative effects on climate change. In this study, the coupled climate feedback response analysis method (CFRAM) is applied to quantify the magnitude of warming over the Himalayas induced by different external forcing factors and climate feedback processes. The results show that snow/ice albedo feedback (SAF) resulted in a warming of approximately 2.6°C and was the primary contributor to enhanced warming over the Himalayas in recent decades. This warming was much greater than the warming induced by dynamic and other radiative factors. In particular, the strong radiative effects of BC on the warming over the Himalayas are identified by comparing control and BC-perturbed experiments of the Community Earth System Model (CESM). As a result of strong BC effects on the Himalayas, evaporation and reduced precipitation were strengthened, accounting for local drying and land degradation, which intensified warming. These results suggest that more investigations on the local radiative effects on the climate and ecosystem are needed, especially in the high-altitude cryosphere.

1. Introduction

Black carbon (BC) aerosol is considered the second most important factor for global climate change due to its high radiative forcing after carbon dioxide (CO₂)

(Jacobson 2001). BC mainly originates from human activities (Bond et al. 2013). Once emitted into the atmosphere, BC can be transported long distances and is removed by wet deposition, such as rain and snow, or by dry deposition (Flanner et al. 2007). The effects of BC on climate change mainly include directly heating atmosphere (e.g., Jacobson 2001; Kopacz et al. 2011; Y. Xu et al. 2016), reducing surface-incident solar radiation (dimming) (e.g., Ramanathan and Carmichael 2008), and reducing snow or ice reflectance after deposition (surface darkening) (e.g., Hansen and Nazarenko 2004;

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-18-0720.s1>.

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Flanner and Zender 2006; He et al. 2014; Lee et al. 2017). The snow/ice-covered regions in the Himalayas and Tibetan Plateau (TP) are more affected by the strong light-absorbing ability of BC than other regions because they are closer to the source of Asian BC emissions and exposed to high levels of sunlight due to their subtropical location (Xu et al. 2009; Lamarque et al. 2010). Both snow/ice surface darkening and the atmospheric heating due to BC generate a forcing that then triggers a positive snow/ice albedo feedback (SAF). The role of the SAF in climate forcing has received increasing attention in recent years, especially for the TP (Flanner et al. 2009; Kopacz et al. 2011; Y. Xu et al. 2016; Jiang et al. 2017). The efficacy of radiative forcing induced by BC deposited onto snow and glaciers is much higher than that by CO₂ because of the strong SAF associated with it (e.g., Hansen and Nazarenko 2004; Flanner et al. 2007; Ye et al. 2012; Doherty et al. 2014). Meanwhile, rapid BC emission growth has occurred over the last few decades in the Asian region, especially in South Asia, and BC deposition in the TP has increased over the past decades (Xu et al. 2009). Modeling studies also showed that the BC transported to the Himalayas and TP increased by approximately 40% from 1996 to 2010, and the highest increase appeared in winter (Lu et al. 2012). BC deposited in snowpack and glaciers has been identified as one of the important radiative factors that accelerated albedo reduction and snow/ice retreat in the Himalayas (e.g., Menon et al. 2010; Y. Xu et al. 2016; He et al. 2018).

As the third largest snow and glacial deposits after those in the Arctic and Antarctic, the melting of snowpack and glaciers over the Himalayas is the source of several major rivers; thus, this area is called the “water tower of Asia” (e.g., Lu et al. 2005; Xu et al. 2008; Palazzi et al. 2013). However, the snow cover (SC) and glaciers over the Himalayas have declined significantly since 1960, which is partly attributed to local radiative effects (Y. Xu et al. 2016; He et al. 2018). In addition to the documented reduction in SC and glaciers (Gardner et al. 2013; Wang et al. 2017), continuous warming has been observed over the Himalayas and TP since 1980, especially in winter (Dimri and Dash 2012; Duan and Xiao 2015; Cai et al. 2017; Ma et al. 2017). Atmospheric BC and deposited BC over the TP have important impacts on regional climate change due to BC-induced effects (e.g., Ramanathan et al. 2007; Meng et al. 2014; Y. Xu et al. 2016). Previous studies focused on the impact of BC on snow/ice and albedo (e.g., Menon et al. 2010; Bond et al. 2013; Y. Xu et al. 2016); however, the role of the SAF triggered by BC on decadal warming over the Himalayas is unclear, and needs to be quantified. In fact, in addition to the trend in snow/ice albedo

variation, changes in CO₂, clouds, atmospheric water vapor, ozone, solar radiation, atmospheric dynamics, land heat storage, and latent and sensible heat fluxes are also possible factors affecting the enhanced warming over the Himalayas in winter (e.g., Liu et al. 2009; You et al. 2010, 2017; Yang et al. 2014, 2018; Duan and Xiao 2015; Jian et al. 2018). Although many studies have explored the roles of individual processes in the enhanced warming over the Himalayas, we will give a comprehensive assessment of the impacts of these factors on surface warming, and focus on the role of SAF triggered by BC in the atmosphere and in surface snow and ice.

Recently, Cai and Lu (2009) proposed an effective method for quantifying spatial variation in individual external forcing factors and climate feedbacks and their contributions to temperature variation. This method has been applied in several studies (e.g., Hu et al. 2016, 2017). As this method can effectively determine radiative effects, we first quantify the contributions of both radiative and dynamic processes to the decadal warming over the Himalayas between two periods: one in the early decades (1981–99) and the other at the end of the enhanced warming event (2000–17). Since the Himalayan region exhibits some of the greatest warming in the TP (Ma et al. 2017), the main objective of this study is to perform a process-based decomposition of the enhanced warming over the Himalayas between the two periods. In particular, the role of BC on SAF over the Himalayas is investigated by the Community Earth System Model (CESM). Meanwhile, the regional changes in the environment over the Himalayas are discussed in the context of decadal warming.

The paper is organized as follows. In section 2, we describe the datasets, methods, and study area used. In section 3, we analyze the variability of albedo and snowpack over the Himalayas, and quantify the relative contributions of SAF and other factors to surface temperature T_s change. We also explore the role of BC as a trigger for the SAF over the Himalayas using the CESM. Discussion and conclusions are presented in sections 4 and 5, respectively.

2. Datasets and methods

a. Datasets

The data from the latest European Center for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim) are used in this study, which are on 37 pressure levels and have a horizontal resolution of $0.75^\circ \times 0.75^\circ$. The variables include downward solar energy flux at the top of the atmosphere (TOA); surface variables, such as net solar radiation, downward

shortwave radiation, surface pressure, snow depth (SD), and sensible and latent heat fluxes; and variables in each layer, such as temperature, cloud amount, cloud liquid/ice water content, specific humidity, and ozone mixing ratio. The albedo used in our study was calculated by the ratio between upward shortwave radiation and downward shortwave radiation at the surface from the ERA-Interim, with upward shortwave radiation obtained by subtracting the downward shortwave radiation from the net solar radiation. As observation stations are sparse and unevenly distributed in the TP, all reanalysis datasets suffer deficiencies, and the uncertainties and biases are much larger than those in the other regions. Note that the effect of aerosols on radiative transfer in the atmosphere is based on climatological aerosols in the ERA-Interim, and the direct signal of aerosol change is not assimilated in the forecast model that produced the ERA-Interim reanalysis (Dee et al. 2011); thus, we are unable to make a direct estimate for the effect of aerosol forcing on the radiative fluxes of the ERA-Interim. However, information on aerosols and other forcing factors is hidden in various observation data due to the presence of various climate feedbacks and climate responses affected by a number of forcing factors. As long as these observations are available for twice-daily assimilation, reanalysis can essentially detect climate trends caused by external forcing (e.g., aerosols or CO₂) even if these forcing factors and their related physical processes are absent from the model used in the data assimilation (e.g., Andersen et al. 2001; Cai and Kalnay 2005; Hu et al. 2017). Previous studies indicated that ERA-Interim is a relatively better choice for analyzing temperature and climate change in the TP (Mao et al. 2010; Wang and Zeng 2012; Shi and Liang 2014; Ma et al. 2017; Peng et al. 2019), where long-term and continuous observations are extremely lacking due to sparse stations and the bitter geographical environment (Wang et al. 2004; Ma et al. 2011). To validate the accuracy of surface albedo change derived from ERA-Interim, the long-term albedo from the satellite-derived climate data record (the Cloud, Albedo, and Radiation dataset from the Advanced Very High Resolution Radiometer sensor) are used as the ground truth for comparison, due to the limitation of ground measurements in spatial coverage and continuity. These albedo data are produced by the Satellite Application Facility on Climate Monitoring (CMSAF) and have high spatial resolution of $0.25^\circ \times 0.25^\circ$ (referred hereafter as CMSAF albedo). Detailed information on these data can be found in Karlsson et al. (2013). The accuracy of CMSAF over the Himalayas was evaluated by Guo et al. (2018), who concluded that CMSAF albedo is suitable for quantifying the long-term albedo change over the Himalayas,

although some values are underestimated compared to the in situ data due to systematic retrieval uncertainty. The monthly T_s during 1980–2016 observed at 10 stations in and near the Himalayas from the China Meteorological Administration is used to verify the temperature change derived from the ERA-Interim, which is obtained from Duan et al. (2018).

The boundary of the Himalayan region is based on the Global Land Cover Network under the Food and Agriculture Organization (FAO) of the United Nations. In addition, the data of anthropogenic BC emissions are derived from the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR5). Because the historical variation in BC emissions during 1850–2000 is given at 10-yr intervals, we use the BC emissions in 2000–10 to represent the present-day distribution of BC emissions over Asia (Bond et al. 2007; Lamarque et al. 2010). The annual-mean BC concentrations from ice cores (in units of ng g^{-1}) from the Noijin Kangsang and Zuoqiupu stations are from Xu et al. (2009). The aridity index (AI), which is calculated using precipitation P divided by potential evapotranspiration (PET), is used to measure the change in terrestrial aridity over the Himalayas. PET was obtained by using the Penman–Monteith algorithm, which is considered the most reliable PET algorithm under all climatic conditions (Dai 2011a; Sheffield et al. 2012). The data used to calculate PET are from the Climate Research Unit (CRU) and the Global Land Data Assimilation System (GLDAS), and the P data are obtained from the Climate Prediction Center (CPC) of the NCEP (Chen et al. 2002). The Palmer drought severity index (PDSI) is a revised drought index used to study global drought risk (Dai 2011b; Wang et al. 2014). The self-calibrated PDSI is used for comparison and verification with the AI over the Himalayas in this study.

b. Study area and decadal change of T_s

The Himalayas located in the southwest of the TP are some of the largest and highest mountains in the world, and this mountain range acts as the barrier between polar and tropical air masses (e.g., Lu et al. 2005; Dimri and Dash 2012). The unique and complex topography, midlatitude location, diverse vegetation types, amplified climate change, and human activities make this region vulnerable to anthropogenic and natural forcing (e.g., Piao et al. 2012; Chen et al. 2014). Moreover, the glaciers and snowpack over the Himalayas provide the headwaters of several major rivers in Asia, which supply freshwater to billions of people (e.g., Lu et al. 2005; Xu et al. 2008). Furthermore, mountain systems are considered a potential region for early detection of global climate change and its associated consequences

FIG. 1. (a) Spatial trend of T_s ($^{\circ}\text{C decade}^{-1}$) over the Himalayas for 1981–2017. (b) The decadal changes in T_s ($^{\circ}\text{C}$) between 2000–17 and 1981–99 over the Himalayas. The asterisk indicates the 95% confidence level. (c) As in (b), but for the sum of changes in T_s due to multiple external forcing factors and climate feedbacks between the two periods, as determined by the CFRAM. The red contour denotes the study area covering the Himalayan region defined by the Global Land Cover Network of the FAO.

on hydrology, ecosystems, society, and economy. However, the Himalayas are one of the least studied regions in terms of climatology and climate projection (Panday et al. 2015; Dimri et al. 2018). Both observations and model simulations of previous studies suggested that consistent warming has occurred over the Himalayas in the present climate, and rapid warming is predicted to affect vegetation cover, plant distribution, and water resources (e.g., Qin et al. 2009; Dolezal et al. 2016; Jin et al. 2018). The increase of temperature was accompanied by retreating snowpack (e.g., Liu et al. 2009; Qin et al. 2009; Rangwala and Miller 2012). Very few studies have comprehensively assessed the enhanced warming over the Himalayas and the role of SAF in the region. Therefore, the study area for the present analysis focuses on the Himalayas in the southern TP (the regions inside the red contour in Fig. 1), within 70° – 100°E and 24° – 36°N with an average altitude of 6000 m.

TABLE 1. Regional average of spatial T_s trends ($^{\circ}\text{C decade}^{-1}$) of seasonal and annual means over the Himalayas for 1981–2017 from ERA-Interim and for 1981–2016 from in situ observations.

Seasonality	Winter	Spring	Summer	Autumn	Annual
Trends from ERA-Interim	0.68	0.34	−0.13	0.01	0.22
Trends averaged over 10 stations	0.77	0.69	0.31	0.52	0.51

To better illustrate the change of seasonality of T_s and the importance of winter warming over the Himalayas, we first calculated the regional averages of spatial T_s trends during 1981–2017 over the Himalayas for the winter (December–February) mean, spring (March–May) mean, summer (June–August) mean, autumn (September–November) mean, and annual mean from the ERA-Interim, which are $0.68^{\circ}\text{C decade}^{-1}$ (winter), $0.34^{\circ}\text{C decade}^{-1}$ (spring), $-0.13^{\circ}\text{C decade}^{-1}$ (summer), $0.01^{\circ}\text{C decade}^{-1}$ (autumn), and $0.22^{\circ}\text{C decade}^{-1}$ (annual), respectively (Table 1). Based on the T_s trends at 10 available stations that mainly distributed in and near the southern Himalayas, the regional-mean T_s trends during 1981–2016 are $0.77^{\circ}\text{C decade}^{-1}$ (winter), $0.69^{\circ}\text{C decade}^{-1}$ (spring), $0.31^{\circ}\text{C decade}^{-1}$ (summer), $0.52^{\circ}\text{C decade}^{-1}$ (autumn), and $0.51^{\circ}\text{C decade}^{-1}$ (annual) (Table 1). The T_s trends from both ERA-Interim and in situ observations indicate that the winter-mean T_s has the most significant increasing trend, contributing most to the annual T_s warming over the Himalayas, which is consistent with previous finding (Du et al. 2004; Chen et al. 2006; Duan and Xiao 2015; Ma et al. 2017). The spatial trend of winter-mean T_s over the Himalayas during 1981–2017 (Fig. 1a) displays significant increasing trends in most regions, indicating an obvious warming during the whole study period. Note also that the spatial T_s had a weak variation before 2000 but an accelerated warming trend after 2000 (Fig. S1 in the online supplemental material), and that the observed T_s averaged from the 10 available stations over the Himalayas exhibited an increasing trend of $0.23^{\circ}\text{C decade}^{-1}$ for 1981–99 and of $0.49^{\circ}\text{C decade}^{-1}$ for 2000–16, suggesting an accelerated warming in the second period. Therefore, we choose 1981–99 as the first period and 2000–17 (with an accelerated warming) as the second period. We calculate the difference in the mean T_s between the two periods (Fig. 1b) using the ERA-Interim data. The distribution of the mean T_s difference between the two periods shows that enhanced warming occurred over the Himalayas (Fig. 1b), which has a highly similar spatial pattern as the trend of T_s during 1981–2017 in winter (Fig. 1a), indicating that the difference between the two periods could roughly represent the net decadal warming over the Himalayas during the whole study period.

c. Coupled climate feedback response analysis method

To quantify the relative contributions of different external forcing factors and climate feedbacks to enhanced warming between the two periods over the Himalayas, we use the coupled climate feedback response analysis method (CFRAM) proposed by Cai and Lu (2009). Based on the local energy balance, the CFRAM considers the entire atmospheric column and the land surface as a coupled surface–atmosphere system. For a given horizontal grid point in the surface layer, the difference in the total energy balance between two climatic states (1981–99 and 2000–17) can be described as follows:

$$\Delta S - \Delta R + (\Delta Q_{\text{LH}} + \Delta Q_{\text{SH}} + \Delta Q_{\text{sfc}} + \Delta Q_{\text{atm}}) = 0, \quad (1)$$

where ΔS and ΔR represent the differences in the absorbed shortwave radiation flux and emitted longwave radiation flux, respectively, between the two periods. Using linear approximation, $\Delta S - \Delta R$ can be divided into partial radiative energy perturbations due to variations in temperature $[(\partial R/\partial T)\Delta T]$, albedo (ΔS_{albedo}), solar radiation (ΔF_s), CO_2 concentration $[\Delta(S - R)_{\text{CO}_2}]$, ozone $[\Delta(S - R)_{\text{O}_3}]$, water vapor $[\Delta(S - R)_{\text{wv}}]$, and clouds $[\Delta(S - R)_{\text{cloud}}]$ between the two periods. This is shown below:

$$\begin{aligned} \Delta S - \Delta R = & -(\partial R/\partial T)\Delta T + \Delta S_{\text{albedo}} + \Delta F_s \\ & + \Delta(S - R)_{\text{CO}_2} + \Delta(S - R)_{\text{O}_3} \\ & + \Delta(S - R)_{\text{wv}} + \Delta(S - R)_{\text{cloud}}, \end{aligned} \quad (2)$$

where $\partial R/\partial T$ is the Planck feedback matrix whose j th column represents the vertical profile of the longwave radiative energy perturbation due to 1-K warming in the j th layer alone. The Fu–Liou radiative transfer model (Fu and Liou 1992) is used to obtain the Planck matrix and radiative energy perturbations in Eq. (2). In addition to these radiative terms, the nonradiative and heat storage terms [the terms in the parentheses in Eq. (1)] also contribute to the total column energy budget, including the energy perturbations mainly due to the surface latent heat flux (ΔQ_{LH}), surface sensible heat flux (ΔQ_{SH}), and land surface energy transport and storage related to rainfall and snow/ice melting (ΔQ_{sfc}), as well as the dynamic processes in the atmosphere (ΔQ_{atm}) (i.e., turbulence, convection, and large-scale atmospheric motions). After substituting Eq. (2) in Eq. (1), the temperature variation ΔT can be attributed to albedo feedback and other individual processes as follows:

$$\begin{aligned} \Delta T = & -(\partial R/\partial T)^{-1} [\Delta S_{\text{albedo}} + \Delta F_s + \Delta(S - R)_{\text{CO}_2} \\ & + \Delta(S - R)_{\text{O}_3} + \Delta(S - R)_{\text{wv}} + \Delta(S - R)_{\text{cloud}} \\ & - \Delta Q_{\text{LH}} - \Delta Q_{\text{SH}} - \Delta Q_{\text{sfc}} - \Delta Q_{\text{atm}}], \end{aligned} \quad (3)$$

$$\begin{aligned} \Delta T = & \Delta T_{\text{albedo}} + \Delta T_s + \Delta T_{\text{CO}_2} + \Delta T_{\text{O}_3} + \Delta T_{\text{wv}} \\ & + \Delta T_{\text{cloud}} - \Delta T_{\text{LH}} - \Delta T_{\text{SH}} - \Delta T_{\text{sfc}} - \Delta T_{\text{atm}}. \end{aligned} \quad (4)$$

The CFRAM has been successfully applied to assess polar warming amplification (Lu and Cai 2010), temperature changes related to El Niño–Southern Oscillation (Deng et al. 2012; Hu et al. 2016), global warming (Hu et al. 2017), and diurnal T_s changes over the TP and southeastern China (Yang and Ren 2017). We apply the CFRAM to evaluate the roles of local radiative factors in the enhanced warming over the Himalayas, and compare their roles with the roles of other dynamic processes.

d. Outputs of climate model experiments

To understand how much BC aerosol forcing derives the SAF over the Himalayas, the outputs of control and BC-perturbed simulations in Xu and Xie (2015) are used in our study. CESM1 was used in these simulation experiments; the overall climate simulation by CESM1 is improved over the previous version (Meehl et al. 2013). These simulations have some main advantages of studying the effects of BC on regional climate change over the Himalayas: The simulated BC forcing constrained by multiple observation sources is close to the observationally constrained values (Xu 2014; Y. Xu et al. 2016). These simulations were conducted using a global model to more accurately account for BC effects on snow and ice, in which the atmospheric BC effect was estimated by a fully coupled ocean–atmosphere–land model at a high resolution of $1^\circ \times 1^\circ$, while the deposited BC effect was calculated in its new land model by incorporating the snow and ice albedo change (SNICAR) module (Lawrence et al. 2011; Xu 2014; Y. Xu et al. 2016). Both atmospheric and deposited BC effects on regional climate change are taken into consideration in this study. Moreover, the subgrid processes in the new land model, including melting, metamorphism, deposition, and redistribution, are considered in SC fraction parameterization (Lawrence et al. 2011), which is versatile in simulating snowpack over the Himalayas. Compared with previous model version, the mean bias of all-sky albedo between new land model of CESM1 and MODIS is reduced throughout the tropics and midlatitudes, and the corresponding mean biases for locations/months with snow cover vary between -5% and 2.9% (Lawrence et al. 2011). In addition, the control

and perturbed experiments include the effects of other major aerosol species such as mineral dust, sulfates, and sea salt on glaciers and snow surfaces, because the aerosol processes are interactively treated in a new modal aerosol module, which is able to simulate the spatial distribution of aerosol optical depth in Asia and compares well with satellite observations (Liu et al. 2012). However, the vast majority of dust/debris on the Himalayas glaciers usually comes from the local environment (Benn et al. 2004). Many glaciers in the Himalayas are covered with rock debris that modifies the local heat transfer and affects the ablation process of snow and ice (Scherler et al. 2011). The thickness and energy balance of the debris layer result in differential surface ablation on the Himalayan glaciers (Rowan et al. 2015). These processes and properties in the complex terrain are hard to represent in CESM and other global models. Thus, our estimates of BC-induced effect on snow/ice may be partially exaggerated, due to the fact that the contribution rate of BC to albedo change varies with surface conditions (e.g., fresh snow, aged snow, and dirty snow) (Qu et al. 2014). Compared with previous coarse-resolution global models (Flanner et al. 2009; Menon et al. 2010; Qian et al. 2011; Jiang et al. 2017), the present global model of CESM is a major step forward in terms of spatial resolution ($1^\circ \times 1^\circ$), which helps better resolve the complex topography in our study region (Ménégoz et al. 2013; Y. Xu et al. 2016).

The ERA-Interim reanalysis does not allow us to obtain a direct estimate of the aerosol forcing effect on radiative fluxes and albedo, so overall radiative effects of atmospheric and deposited BCs on regional climate change are investigated by analyzing the responses of related climate variables due to BC forcing in the simulation. The control experiment was run for 319 years to avoid any discernible drift in the mean climate state, and an additional 75 years for analysis. Based on the control experiment at the end of the 319th year, the perturbed simulation was run by instantaneously increasing BC emissions to the present-day level and maintaining it for 75 years. The first 15 years of the 75 years are considered the spinup period and the impact of natural decadal variability is eliminated by long-term averaging. The differences in 60-yr-mean albedo, net solar flux, SD, and T_s between the BC-perturbed and control simulations are calculated, which denote the radiative effects due to increased BC. Although the model estimate of radiative forcing by BC has some uncertainties due to both BC-snow mixing state and snow grain shape that determine the strength of snow/ice albedo effect (Liou et al. 2014; He et al. 2017), these model simulation outputs have been used in various climate studies, such as the studies on the ocean mediation of responses to aerosols (Xu and Xie 2015), high-altitude snow retreat in the TP (Y. Xu et al. 2016), aridity

FIG. 2. (a) Spatial distribution of BC emission ($10^{-10} \text{ kg m}^{-2} \text{ s}^{-1}$) in the 2000s over Asia. The gray contour denotes the elevation of 1800 m covering the TP, and the blue contour represents the Himalayan region. Locations 1 and 2 marked with black dots are Noijin Kangsang and Zuoqiupu stations. (b) Annual mean BC concentration (ng g^{-1}) in the Noijin Kangsang ice core (29.04°N , 90.02°E) from 1980 to 2006. (c) As in (b), but for the Zuoqiupu ice core (29.21°N , 96.92°E).

analysis (Lin et al. 2016), and dryland warming over East Asia (Zhang et al. 2017). More detailed information on these model simulations can be found in Xu and Xie (2015).

3. Results

a. Observed BC over the Himalayas

Figure 2a shows the spatial distributions of BC emissions in Asia. The Himalayas located in the southwest of the TP are in the middle of two major BC source regions, South Asia and East Asia. Several modeling studies also found that South Asia accounted for 67% of the BC

FIG. 3. (a) Decadal differences in albedo (%) in winter over the Himalayas between 2000–17 and 1981–99. The asterisk indicates the 95% confidence level. (b) Time series of albedo anomalies (%) in winter over the Himalayas. The red and blue lines are based on the ERA-Interim and CMSAF albedo, respectively. The solid lines represent the time series of the original data, and the dashed ones are the 5-yr running mean.

transported to the Himalayas and TP (Ramanathan and Carmichael 2008; Menon et al. 2010; Zhang et al. 2015), and the increased BC emissions from South Asia during the past decades caused more BC in the Himalayas, especially in winter (Ming et al. 2008; Ramanathan and Carmichael 2008; Menon et al. 2010; Lu et al. 2012). Moreover, previous studies found that two pathways contributed to the primary transport of BC to the Himalayas: Long-distance and high-elevation transport has been related to westerlies and synoptic systems under dry winter monsoon conditions (Ming et al. 2008), while short-distance and low-elevation transport is affected by the Asian summer monsoon (e.g., Jin et al. 2015; Zhang et al. 2015; Pu and Ginoux 2016). Figures 2b and 2c show the trends of annual-mean BC concentrations from ice cores collected at the two stations over the Himalayas, both of which exhibited significant increasing trends from 1980 to 2006 but had regional differences in terms of increasing rate. The increased amplitude of normalized BC concentration was $5\% \text{ yr}^{-1}$ ($10\% \text{ yr}^{-1}$) with significance at the 90% (99%) confidence level at Noijin Kangsang (Zuoqiupu) station in the past three decades, according to a two-tailed

Student's *t* test. There was evidence indicating that BC emissions increased rapidly since the 1990s in South Asia, and the BC deposition continued to increase over the past 60 years at multiple stations in the Himalayas (Xu et al. 2009; Ji et al. 2015). The results based on Fig. 2 and previous studies (Ming et al. 2008; Xu et al. 2009; Lu et al. 2012; Qian et al. 2015) suggest that BC received by the Himalayas exhibits an increasing trend over the past decades, which may affect local radiative processes. For instance, previous studies concluded that BC and dust in the surface snow/ice contributed to about 20% of the albedo reduction in the TP (Qu et al. 2014; Li et al. 2017; Schmale et al. 2017), and BC on the surface of the snow induced an increase in temperature of 0.1° – 1.5°C and a decrease in snow water equivalent of over 25 mm in the western TP and Himalayas (Ji 2016).

b. Reductions in albedo and snowpack over the Himalayas

The winter albedo exhibits a significant decrease over the Himalayas in the recent period (i.e., 2000–17) compared to the previous period (i.e., 1981–99) (Fig. 3a), and the decreasing centers are distributed in the west of the

FIG. 4. (a) Decadal difference (%) in SC in winter over the Himalayas between the two periods. The asterisk indicates the 95% confidence level. (b) Decadal mean of the average SD anomalies (cm) in winter over the Himalayas. “Mean1” represents the SD anomaly in the earlier decades of 1981–99, “mean2” represents the SD anomaly in the later decades of 2000–17, and “difference” is the difference between the two periods.

Ngari Prefecture and upstream of the Yarlung Zangbo River (Fig. 3a), corresponding to the enhanced warming centers over the Himalayas (Fig. 1b). This negative pattern and change range of albedo from the ERA-Interim are consistent with the findings of other studies (Ming et al. 2015; Chen et al. 2017). To validate the accuracy of albedo change using the ERA-Interim, Fig. 3b shows the time series of albedo anomalies for ERA-Interim and CMSAF in winter over the Himalayas during 1981–2017. We can see that the albedo anomalies and trends from the ERA-Interim are broadly similar to those from the CMSAF. The correlation coefficient R between the two time series is 0.54 ($P > 95\%$), which is 0.78 ($P > 95\%$) for the 5-yr running mean (Fig. 3b). Both ERA-Interim and CMSAF albedo exhibit decadal decrease but the reduction in ERA-Interim albedo is greater. Overestimation of the decadal decrease in ERA-Interim albedo is mainly caused by the obvious biases during 2004–10, which might be related to the uncertainties of various sources from assimilation system used for the ERA-Interim and satellite retrieval error in the CMSAF (Guo et al. 2018). Meanwhile, the change from ERA-Interim albedo over the Himalayas is essentially in agreement with that of SC from the National Snow and Ice Data Center (figure not shown); their R is 0.72 ($P > 95\%$). These results indicate that ERA-Interim albedo over the Himalayas is reliable and valuable. The qualitative changes of albedo on the decadal scale from the ERA-Interim are consistent with those from satellite retrieval data. However, some values are overestimated due to larger data uncertainties over the Himalayas than over other regions. This calls for more attention to be paid to the improvement of data quality in future studies, and

highlights an urgent need for extensive in situ measurements in the region.

Based on ERA-Interim, the albedo anomaly presents an overall decreasing trend but a weak change from 1981 to 1996. Then, the albedo decreased remarkably from 1998 until 2008 and increased slightly after 2008. Both albedo (Fig. 3b) and T_s (Fig. S1) exhibited slight variations before 2000; however, a significant decreasing trend in albedo and an accelerated warming trend in T_s appear in the recent period. Such consistent variability between albedo and T_s suggests that a radiative effect is obvious over the Himalayas, and the decadal albedo reduction and warming are partly attributed to the light-absorbing constituents, such as BC and dust (e.g., Ming et al. 2008; Lau et al. 2010; He et al. 2018). We examine the variability of snowpack over the Himalayas between the two periods, and calculate the decadal change in winter mean SC over the Himalayas between the two periods (Fig. 4a). Figure 4a shows a decrease of SC over the Himalayas in the recent decades in contrast to the previous decades, which is consistent with the change in albedo (Fig. 3a). Figure 4b shows the decadal mean of SD and differences in SD anomalies between the two periods over the Himalayas. The mean SD anomalies in the periods of 1981–99 and 2001–17 are 0.31 and -0.35 cm, respectively, and the SD anomaly difference between the two periods decreases by 0.66 cm. This result suggests that similar to albedo, SD decreased more obviously in the recent decades than in the previous decades. Meanwhile, the most rapid and steady reduction of glaciers based on observations occurred in the Himalayan region (Gardner et al. 2013). Figures 3 and 4 illustrate that the reductions in SC and SD are accompanied by the reduction in albedo, and then cause

FIG. 5. Partial changes in T_s ($^{\circ}\text{C}$) due to individual external forcing factor and climate feedback between 2000–17 and 1981–99 over the Himalayas: (a) T^{albedo} ; (b) T^{SH} ; (c) T^{LH} ; (d) T^{sfc} ; (e) T^{cloud} ; (f) T^{atm} ; (g) T^{CO_2} ; (h) T^{wv} ; and (i) T^{O_3} .

positive SAF, which may have an important impact on the enhanced warming over the Himalayas. However, the quantification of the contributions due to SAF and its warming mechanism over the Himalayas are still unclear. We will quantify the relative contributions of different factors to temperature variation in Fig. 5.

c. The key role of SAF in the enhanced warming

We apply the CFRAM to decompose the observed total temperature change in winter between the two periods into nine partial changes related to individual radiative and dynamic processes (Fig. 5). The sum of all partial temperature changes shown in Fig. 1c is highly consistent with the observed temperature but slightly smaller (Fig. 1b). Small differences between the sum of partial temperature changes derived from the CFRAM and original temperature change directly obtained from ERA-Interim are mainly due to the errors in linearization of the radiative transfer model, which are close to zero and much smaller than individual temperature changes (Hu et al. 2017). These results indicate that the CFRAM is useful for quantitative attribution of individual processes in T_s changes between the two periods. Figure 5a suggests that SAF primarily contributes to the strong surface warming over the Himalayas, with mean warming of 2.6°C . The weakened evaporation feedback accompanied by decreased upward surface latent heat flux in the recent decades is the second largest contributor to enhanced

warming over the Himalayas, as shown in Fig. 5c. However, in winter the downward sensible heat flux is one of the major sources of heat to the TP surface. Compared with the previous decades, the warmer temperatures in the recent decades in the Himalayas indicate a decreased air-surface temperature difference and reduced downward sensible heat flux, resulting in surface cooling (Fig. 5b). The sensible-heat-related surface cooling and latent-heat-related surface warming (Fig. 5c) nearly offset each other over the Himalayas. In addition, the land storage has a warming effect to the west of the Himalayas but a cooling effect in the other areas of the Himalayas (Fig. 5d). Only a small area with cooling induced by cloud feedback appears in the western Himalayas (Fig. 5e), which offsets the warming due to land storage change in the same region (Fig. 5d). The slight cooling due to atmospheric dynamic change appears in the Himalayan region (Fig. 5f), which is smaller than that caused by the SAF. However, the partial temperature changes due to changes in CO_2 (Fig. 5g), water vapor (Fig. 5h), and ozone (Fig. 5i) are relatively weak and negligible compared to the changes due to the other processes over the Himalayas during winter. Therefore, these results illustrate that the SAF is the primary contributor to the enhanced warming in winter over the Himalayas, and the effects due to surface sensible and latent heat fluxes counteract, while the temperature changes driven by the remaining processes are much

FIG. 6. Changes in response to BC forcing between perturbed and control simulations: (a) albedo (%), (b) clear-sky net solar flux at surface (W m^{-2}), (c) SD (cm), and (d) T_s ($^{\circ}\text{C}$). The cross indicates the 95% confidence level.

weaker. Using albedo and other datasets from the Modern-Era Retrospective Analysis for Research and Applications (MERRA) and JRA-55 reanalysis and surface energy budget equation, Su et al. (2017) suggested that positive surface albedo feedback dominated the warming in winter over the TP, which is consistent with our results derived from the ERA-Interim. Moreover, Guo et al. (2018) showed that the large negative values of SAF over the Himalayas derived from the ERA-Interim are similar with those of satellite-based datasets during 1982–2005.

d. The role of BC in SAF over the Himalayas revealed by model

As noted earlier, the enhanced warming over the Himalayas was dominated by the SAF process in winter during the recent decades. As a typical radiative factor, the observed BC concentration exhibited an increasing trend over the Himalayas in the recent decades (e.g., Lu et al. 2011; Zhang et al. 2015; Jiang et al. 2017). The effects of surface darkening and atmospheric heating due to BC cause snow melting and surface warming (e.g., Jacobson 2001; Hansen and Nazarenko 2004; Kopacz et al. 2011; Y. Xu et al. 2016), which further

amplify the original SAF. However, previous studies showed that the positive radiative forcing at the surface induced by surface darkening effect of BC is even larger than that forced by atmospheric BC effect (e.g., Menon et al. 2010; Y. Xu et al. 2016; Jiang et al. 2017). The role of BC in the SAF process over the Himalayas is investigated by the CESM; in particular, the BC emissions in the model were increased instantaneously from the preindustrial level (used in the control run) to present-day level in the perturbed run. The model outputs help understand the responses of albedo and snowpack to BC, although they are not a reproduction of the actual physical process during the twentieth century. The changes in the albedo and clear-sky net solar flux over the last 60 years in the Himalayas between BC-perturbed and control experiments are shown in Figs. 6a and 6b, respectively. The response of albedo to BC declines by 4.08% in the Himalayas, and exhibits a highly similar pattern to that of observed albedo (Fig. 3a and Table 2). The clear-sky net shortwave flux at the surface due to BC exhibits an increase of up to 10.13 W m^{-2} , implying that an obvious decrease in albedo due to BC accompanies by a considerable increase in surface absorption of solar

TABLE 2. Changes in albedo, SAT, P , PET, AI, wind speed u_{10} , and RH averaged over the Himalayas due to BC forcing.

Simulations	Albedo (%)	SAT ($^{\circ}\text{C}$)	P (mm day^{-1})	PET (mm day^{-1})	AI	u_{10} (10^{-2} m s^{-1})	RH (%)
BC forcing	-4.08 ↓	+1.12 ↑	-0.03 ↓	+0.09 ↑	-0.33 ↓	-9.77 ↓	-2.57 ↓

FIG. 7. As in Fig. 6, but for (a) P (mm day^{-1}) and (b) PET (mm day^{-1}) due to BC. The cross indicates the 95% confidence level.

radiation in the Himalayas (Fig. 6b), which may accelerate regional warming. Moreover, the large-scale pattern of decreased albedo and increased solar flux due to BC (Figs. 6a,b) corresponds to enhanced warming due to the SAF over the Himalayas, as shown in Fig. 5a, suggesting that BC plays an essential role in reducing albedo and enhancing warming over the Himalayas.

Figure 6c displays the variable related to the snowpack in response to BC in the same model, and the response of SD to BC exhibits a large decrease by up to 1.4 cm on average throughout the Himalayas. The pattern of snow retreat (Fig. 6c) is similar to that of observed snowpack (Fig. 4) and of albedo (Fig. 6a). Moreover, the T_s in response to BC (Fig. 6d) displays significant warming due to BC-induced increase in clear-sky net solar flux (Fig. 6b) and reductions in albedo (Fig. 6a) and SD (Fig. 6c). Figures 5 and 6 imply that BC forcing results in a decrease in albedo, leading to increases in surface net solar flux and snow retreat and contributing to the enhanced warming through amplifying the original SAF over the Himalayas. We confirm that the SAF amplified by BC plays a key role in the enhanced warming over the Himalayas during winter, when compared with other multiple radiative and dynamic processes considered by the CFRAM (Figs. 5 and 6). As discussed in many previous studies (Liou et al. 2014; He et al. 2017), the model uncertainties for simulated BC effect on snow/ice may originate from snow grain shape, snow aging, snow/ice surface conditions, BC-snow mixing state, and the mixing of BC with other components (e.g., dust and debris). Despite these uncertainties, our results provide the reference information that BC forcing amplifies the local SAF over the Himalayas, which needs validation by using more observations and simulations. In addition, changes of other light-absorbing aerosols such as dust and organics may also induce reduction of snow/ice and albedo (Qu et al. 2014; Wang et al. 2015; Lau and Kim 2018; Yao et al. 2019), and more investigations for local radiative

effects and quantifications of different light-absorbing factors on regional SAF are urgently needed in the future.

e. Other decadal climate effects induced by BC in the model

In addition to the enhanced warming related to BC, the responses of P and PET and other parameters to BC also vary significantly over the Himalayas. Previous studies showed that anthropogenic BC has an impact on global climate change, especially in Asia (Bauer et al. 2010; Qian et al. 2011; Zhang et al. 2012; Lin et al. 2016; Jiang et al. 2017; Lau and Kim 2018). For instance, Jiang et al. (2017) indicated that the East Asian winter monsoon is intensified by anthropogenic BC-induced TP warming in their simulation, which is characterized by a significant surface cooling in northern East Asia and an acceleration of the jet stream around 40°N. Lin et al. (2016) showed that BC induced a global drying of $1.9\% \text{ }^{\circ}\text{C}^{-1}$, and surface air temperature (SAT) and available energy changes dominated BC-induced PET changes. To investigate the responses of P and PET to BC over the Himalayas, we calculated the difference in long-term-mean P and PET between the control and BC-perturbed simulations over the Himalayas (Fig. 7). The pattern of the P response to BC in Fig. 7a shows a weak decrease in the Himalayas. In contrast to the P change, the PET response to BC (Fig. 7b) exhibits an evident enhancement over the Himalayas. The magnitude of increased PET is significantly greater than that of reduced P , indicating that the change in PET response is more obvious than the change in P due to BC over the Himalayas. The SAT, relative humidity (RH), and wind speed affect the change in PET (Byrne and O’Gorman 2013; Lin et al. 2015). As listed in Table 2, the SAT due to BC increases by 1.12°C , while the wind speed and RH due to BC decrease by 9.77 (in units of 10^{-2} m s^{-1}) and 2.57 (in units of %), respectively. The response of PET to BC shown in Fig. 7b exhibits a feature similar to the temperature change (Fig. 6d and Table 2). This result indicates that temperature is the major contributor of BC-induced PET change over the

TABLE 3. The observed differences in the AI and PDSI between 2000–08 and 1991–99.

Observations	AI	PDSI
Decadal difference	−0.21 ↓	−1.83 ↓

Himalayas, while the effects of wind speed and RH may be relatively small for BC-induced PET change. As illustrated earlier, BC induced surface warming by amplifying the SAF (Fig. 6), which could also increase the PET in the same region (Fig. 7b).

4. Discussion

Temperature, P , wind speed, and RH serve as important climate variables that significantly affect many other ecological and hydrological variables, such as PET; the regional AI, which is regarded as a measure of the dryness of the climate; the PDSI, which is used to study aridity changes; and the normalized difference vegetation index (NDVI), which is considered a good indicator of vegetation cover and productivity. Table 3 lists the observed changes in the AI and PDSI between the two periods over the Himalayas. Both the AI, which is defined as the ratio of P to PET, and the PDSI decreased in the recent decades compared to those in the previous decades, implying that the observed climate over the Himalayas became drier during the recent years (Table 3). Similarly, BC forcing led to a -0.33 decrease in the AI based on regionally averaged P and PET and a 1.12°C increase in the regionally averaged temperature (Table 2), indicating that BC contributed to the drier and warmer climate over the Himalayas. This is in good agreement with observations (Guan et al. 2017). Gao et al. (2015) suggested that the drier and warmer climate in the Himalayas might aggravate land degradation. Meanwhile, both climate change and anthropogenic activities affected vegetation dynamics, and there was an overall increase in vegetation growth over the Himalayas based on annual-mean NDVI and other multiple datasets of long-term observations during 1982–2013 (Piao et al. 2012; Cai et al. 2015; Yao et al. 2019), which was mainly attributed to climate change (Piao et al. 2012; Chen et al. 2014; Cai et al. 2015; Liang et al. 2017). However, significant decreasing trend of annual-mean NDVI appeared in some areas of the Himalayas during 2000–11 (Zhang et al. 2013; Chen et al. 2014), which was affected by decreased P , enhanced drought, and increased anthropogenic activities in the recent years through reducing soil water content and water availability for plant growth and thus weakening or offsetting the positive effect of warming on NDVI (Zhang et al. 2013; Chen et al. 2014; Piao et al. 2014; Cai

et al. 2015; Liang et al. 2017). These findings suggest that the warmer and drier climate led to dry conditions over the Himalayas in the recent period, which might be unfavorable for vegetation growth and aggravate land degradation (Zhang et al. 2013; Chen et al. 2014).

In addition to the changes in PET, AI, PDSI, and NDVI, the warming of the climate over the TP also induced changes in atmospheric circulation, such as driving frequent haze events in eastern China (X. Xu et al. 2016), suggesting that the enhanced warming over the TP had a significant impact on the quality of people's lives. Figure 8 lists possible factors and processes of climate change that affect the regional vegetation and environment. In detail, BC induced enhanced warming by amplifying the SAF and resulted in changes in P , wind speed, and RH, which caused the warmer and lower moisture conditions over the Himalayas and then contributed to increases in PET and regional drying and a decrease in vegetation, aggravating regional desertification and land degradation, which conversely intensified local warming and affected climate change (Huang et al. 2016). Therefore, changes in climate variables such as temperature, P , wind speed, and RH may alter regional vegetation, environment, and water storage (Huang et al. 2018). Our results suggest that the implications of enhanced warming dominated by the SAF, which affects the regional vegetation and environment, should be considered when creating environmental protection policies over the Himalayas and other similarly vulnerable areas.

5. Conclusions

The Himalayas are always covered by snow, which makes this region an ideal area to detect the radiative effect of snow. During the process of determining the effect of snow on temperature change, it was found that a large amount of BC is transported from South Asia to the Himalayas (e.g., Ming et al. 2008; Xu et al. 2009; Lu et al. 2012; Qian et al. 2015). With the increased BC aerosol over the Himalayas, we found that albedo and snowpack obviously decreased during the recent period, and these decreases corresponded to the enhanced warming over the past decades in the region. Then, we analyzed the role of SAF triggered by BC in the enhanced warming over the Himalayas, as the Himalayas are considered one of the few areas at the midlatitudes of the Northern Hemisphere that exhibited accelerated warming during the recent period, especially in winter. Furthermore, we provided an assessment of the relative contributions from multiple external forcing factors and climate feedback processes to the enhanced warming over the Himalayas.

FIG. 8. Schematic diagram of the impact of climate factors on the environment.

Our results indicate that the BC-amplified SAF dominated the enhanced warming in winter over the Himalayas during the recent decades. The variations in temperature due to changes in land energy transport/storage, clouds, dynamic processes in the atmosphere, water vapor, CO₂, and ozone were relatively weak. Simulation results reveal that BC amplified the SAF by increasing clear-sky net solar flux and decreasing albedo and snowpack, which was conducive to the enhanced warming over the Himalayas. In addition, this local SAF amplified by BC also contributed to slightly reduced *P* (drying) and changes in wind speed and RH over the Himalayas. As important climate factors, enhanced warming and reduced *P* lead to increased PET and drying, which will exacerbate local desertification and land degradation. Conversely, these conditions are likely to amplify the local warming over the Himalayas. Therefore, the impact of climate change related to BC on the regional vegetation and environment over the Himalayas requires further investigation.

Acknowledgments. This study was supported by the National Natural Science Foundation of China (91637312, 41722502, 41575006, and 41521004) and the China 111 Project (B 13045). The authors thank the ECMWF, FAO, CMSAF, IPCC AR5, Earth System Research Laboratory, CRU, GLDAS, and the CPC for

providing accessible and excellent datasets that made this study possible. The ERA-Interim data are available for download (<http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/>). The data of BC emissions are downloaded from <http://tntcat.iiasa.ac.at:8787/RcpDb/dsd?Action=htmlpage&page>. The CO₂ concentration data are downloaded from the Earth System Research Laboratory (<http://www.esrl.noaa.gov/gmd/ccgg/trends/>). The weekly SC data are from the National Snow and Ice Data Center (<https://nsidc.org/data/>), and the monthly SC data were converted from the weekly data in Wang et al. (2017). Dai's self-calibrated PDSI data are downloaded from <http://www.cgd.ucar.edu/cas/adai/data-dai.html>. The authors thank Yangyang Xu for providing the BC-perturbed and control simulation outputs, and thank Baiqing Xu and Cenlin He for providing the BC concentration data in the Himalayas.

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