Process-based attribution of long-term surface warming over the Tibetan Plateau

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Abstract
The Tibetan Plateau (TP) has experienced a rapid warming since the 1980s even during the global warming hiatus. Using the European Centre for Medium-Range Weather Forecasts Interim (ERA-Interim) reanalysis and a climate feedback-response analysis method, we attribute the long-term change in near-surface air temperature (SAT) over the TP to external forcing and internal climate processes. Results indicate that over the eastern and central TP, a reduction of surface sensible heat flux going from the surface to the atmosphere and the moistening of the atmosphere are responsible for the increase in the annual mean SAT. In the western TP, however, the most important contributors to surface warming are surface latent heat flux, surface albedo, and clouds. In summer (June–July–August), strong water vapour convergence with anomalous ascending motion results in more cloud cover over the central and eastern TP, causing a stronger cooling effect at the surface which is compensated by changes in water vapour and surface sensible heat flux. In winter (December–January–February), the change in surface albedo shows a significant increasing trend in the past 40 years, showing less net downward solar radiation to warm the surface. However, the decadal reduction of surface sensible heat flux and increase of heat storage result in a warming surface.

KEYWORDS
climate feedback-response analysis method, seasonality, surface warming, Tibetan Plateau

1 | INTRODUCTION

The Tibetan Plateau (TP) is one of the highest and largest plateaus in the world, with an area of 2.5 million square kilometres and an average altitude of more than 4,000 m. The dynamic and thermodynamic forcing of the TP can influence regional and global climate significantly (Wang et al., 2008; 2019; Bao et al., 2010; Wu et al., 2015; Lu et al., 2017). Meanwhile, as a semi-arid region, the TP is one of the most sensitive areas to climate change (Liu and Chen, 2000; Yao et al., 2000; Guan et al., 2019). Therefore, it is important to understand the change in TP climate for the sake of both science advancement and societal wellbeing. Previous studies on the long-term change in TP near-surface air temperature (SAT) have reached a consensus: the TP has experienced a rapid
warming in the last few decades (Wang et al., 2008; Li et al., 2010; Duan and Xiao, 2015). However, since observational data are limited over the TP, especially in the western TP, there exists a large uncertainty about the rate of warming over TP. Liu and Chen (2000) utilized limited observational data to show that during the period of 1955–1996, the change in SAT over the TP exhibited a positive trend of 0.16 K per decade. Xu et al. (2017) found that the SAT over the TP experienced a warming rate of 0.30 K per decade in arithmetic average and 0.35 K per decade in area-weighted average during the period of 1961–2015. Furthermore, there were seasonal difference on the long-term trend of SAT over the TP. Zou et al. (2014) showed that the dramatic warming trends over the TP were 0.33 K per decade in the annual mean, with a relatively weak warming trend in summer (0.22 K per decade) but pronounced warming in winter (0.47 K per decade), based on the SAT in the European Centre for Medium-Range Weather Forecasts Interim (ERA-Interim) dataset during the period of 1979–2010. However, Duan and Xiao (2015) analysed station data and found that the warming trends over the TP were comparable with that over the rest of China in spring and autumn, but the TP had larger warming trends than the rest of China in summer and winter. Nevertheless, the temperature change in summer and winter has received more extensive attentions. Moreover, not only the mean SAT, but also the maximum and minimum SAT showed significant increasing trends (Liu et al., 2009; 2018; You et al., 2016). Besides, the warming over TP is non-homogeneous in terms of spatial distribution, with stronger warming trend in the higher elevations than in the lower elevations (Yan et al., 2016; An et al., 2017). In general, the TP is one of the most significant warming regions over the world (Duan and Wu, 2006).

The possible mechanism for the warming over the TP has been discussed extensively. Changes in surface snow and ice cover were considered as the leading cause of the TP warming. Wang et al. (2018) found that there was a remarkable decreasing trend for snow cover in the western and southern TP, but a significant increasing trend in the central-eastern region for snow cover. The warming surface over the TP would make the snowline retracted and the surface albedo decreased, resulting in larger net downward solar radiation to further warm the surface (Ghatak et al., 2014; You et al., 2016). Cloud may be another factor accounting for surface warming over the TP. Duan and Wu (2006) found that low-level cloud exhibited a prominent increasing trend at night, which would likely cause a warming surface at night due to stronger downward longwave radiation. In addition, the reductions of low-level cloud and total cloud in daytime accompanied by an increase in surface net solar radiation also contribute to a warmer surface over the TP. Rangwala et al. (2009) indicated that the increased water vapour over the TP in winter would trap more longwave radiation in the lower-troposphere to warm the surface air. According to station data, the specific humidity at the surface layer showed an increasing trend during 1984–2006 (Yang et al., 2014). The enhanced anthropogenic greenhouse gas effect also increases SAT over the TP (Duan et al., 2006). In addition, Su et al. (2017) and Gao et al. (2019) quantitatively analysed the warming amplification over the TP by applying the energy budget equation, and revealed that the decrease in surface albedo and increase in longwave radiation related to CO₂ concentration, water vapour, and air temperature contributed to a large portion of surface warming but less surface warming was attributed to the changes in surface heat fluxes and surface shortwave radiation related to cloud. However, the relative contributions of individual processes were not fully clear and a more detailed discussion needs to be carried out. In this study, we perform a comprehensive quantitative analysis to discuss the relative contributions of dynamic and thermodynamic processes to the surface warming over the TP in the recent decades.

A well-documented method named Coupled surface-air Feedback Response Analysis Method (CFRAM; Cai and Lu, 2009; Lu and Cai, 2009a) is designed to quantitatively attribute the global or regional temperature change to individual external forcing (e.g., CO₂ concentration, ozone, and solar radiation) and internal feedback processes including surface albedo, cloud, water vapour, surface sensible and latent heat fluxes, land/ocean heat content, and dynamic processes. The CFRAM has been used for revealing the relative contributions of various climate feedback processes to global warming and hiatus (Cai and Tung, 2012; Hu et al., 2017b; 2018b), polar warming amplification (Lu and Cai, 2009b; 2009c), El Nino-related sea surface warming (Park et al., 2012; Hu et al., 2017a), and the temperature bias in climate models (Park et al., 2013; Liu et al., 2015; Yang et al., 2015). Recently, it has also been applied to attribute climate change over the TP, such as decadal changes in diurnal SAT range (Yang and Ren, 2017) and cold bias in model simulation over the TP (Chen et al., 2017b).

In this study, we perform a comprehensive quantitative analysis of TP warming by applying the CFRAM to identify the main processes responsible for the long-term annual mean surface warming over the TP as well as the surface warming in summer and winter in the past few decades. The remaining parts of this paper are organized as follows. Section 2 describes the data and presents the key features of the long-term changes in the annual mean SAT over the TP, as well as changes in SAT in summer and winter. Section 3 describes the method for the process-based
decomposition. The main processes responsible for surface warming are discussed in Section 4. Conclusions and discussion are presented in Section 5.

2 | DATA AND SAT CHANGE OVER THE TP

2.1 | Data

The data used in this study are mainly the monthly mean data obtained from the ERA-Interim reanalysis (Dee et al., 2011), which covers the period from 1979 to 2016, with a horizontal resolution of 1° × 1° and 37 pressure levels ranging from 1,000 to 1 hPa. The variables analysed include U wind component, V wind component, vertical velocity, air temperature, specific humidity, cloud ice and liquid water content, cloud cover, ozone mixing ratio, SAT, upward and downward shortwave radiation, solar radiation at the top of the atmosphere, surface pressure, surface sensible heat flux, and surface latent heat flux. The CO₂ concentration, also from 1979 to 2016, is obtained from NOAA’s Earth System Research Laboratory website (https://www.esrl.noaa.gov/gmd/ccgg/trends/).

Previous studies have examined the quality of the ERA-Interim reanalysis dataset over the TP. It was found that the ERA-Interim provides the most reliable monthly SAT data compared to the in situ observations at 63 Chinese Meteorological Administration stations over the TP (Wang and Zeng, 2012). Bao and Zhang (2013) used the data obtained from the Tibetan Plateau Experiment (TIPEX) as a reference and pointed out that the ERA-Interim had a smaller root mean square (RMS) error and bias than the NCEP_NCAR and the ERA-40. Furthermore, the ERA-Interim provided more reliable surface heat flux in the eastern TP compared with the JRA-55 and the MERRA (Xie et al., 2019), which is consistent with the result of Shi and Liang (2014), who showed that compared with the CFSR, the MERRA, and the JRA-25, the ERA-Interim had the lowest RMS errors for surface heat flux. Lately, Zhao et al. (2018) concluded that the ERA-Interim had the most reliable heat flux according to the observation data obtained from the Third Tibetan Plateau Atmospheric Scientific Experiment (TIPEX-III). These results give us confidence to use the ERA-Interim reanalysis data to describe the climate change over the TP.

2.2 | SAT warming over the TP

The key features of changes in annual, June–July–August (JJA), and December–January–February (DJF) mean SAT over the TP from 1979 to 2016 are shown in Figure 1. The time series of annual mean SAT over the TP exhibits a warming trend of 0.26 K per decade (Figure 1a), which is comparable with the 0.33°C per decade by Zou et al. (2014) but is smaller than 0.44°C per decade based on observation station data (Duan and Xiao, 2015) and larger than the 0.1°C per decade from Yang and Ren (2017) based on a shorter reanalysis record (1979–2013). The large spread of the warming rate over the TP is because that the warming rate is sensitive to analysis period and domain and to data sources. The warming trends in summer and winter are 0.21 and 0.22 K per decade, respectively (Figure 1b,c).

We use the Mann–Kendall test, which is widely used to detect the mutation of a time series such as the abrupt change in mean value to examine whether the change in the SAT over the TP is significant. Figure 1d,e show that the annual and summer mean SATs exhibited an abrupt change around 1998, and the abrupt change point in winter is not as significant as that in summer (Figure 1f). Therefore, the whole period of 1979–2016 is split into two climatic stages: 1979–1997 and 1998–2016. The difference in SAT over the TP between these two periods, denoted by the symbol ‘Δ’, presents a net warming in the past four decades, about 0.62 K in annual mean, 0.47 K in JJA mean, and 0.58 K in DJF mean (Figure 1a–c). The bottom panel shows the spatial distributions of the trends of annual (Figure 1g), JJA (Figure 1h), and DJF (Figure 1i) mean SAT over the TP. The hatched areas indicate the significant warming trends at the 95% confidence level according to two-tailed Student’s t test. The warming trends of annual mean are significant over almost the entire TP. In summer, the warming trends are larger than that in annual mean over the central and northern regions, but not obvious over the southern flank. The warming trends in winter are quite different from the annual mean and JJA mean, and the significant warming areas are in the southern slope and a small part in the central and northwestern TP. Generally, there is evident warming over the entire TP except for a small area in the western TP in the annual mean, as well as in JJA and DJF means.

3 | ANALYSIS METHOD

Based on the total energy balance within an atmosphere-surface column, Lu and Cai (2009a) and Cai and Lu (2009) developed the CFRAM to quantify the addable contributions of external forcing and climate feedback processes to total temperature change. The CFRAM has been adopted to attribute global (Hu et al., 2018a) and regional surface warming (Chen et al., 2017a; Li et al., 2017). Take the difference (Δ) between 1979–1997 and 1998–2016, namely:
\[
\Delta S - \Delta R + \Delta Q_{\text{non-radiative}} = 0 \tag{1}
\]

where \( S \) and \( R \) denote the vertical profiles of solar energy absorbed and the net longwave radiation emitted within each individual layer, respectively. By neglecting the nonlinear interactions among various radiative feedback processes, \( \Delta S \) and \( \Delta R \) can be linearly decomposed as the sum of partial energy perturbations due to individual radiative processes, that is:

\[
\begin{align*}
\Delta S &\approx \Delta S_{\text{WV}} + \Delta S_{\text{CLD}} + \Delta S_{\text{CO}_2} + \Delta S_{\alpha} \\
\Delta R &\approx \Delta R_{\text{WV}} + \Delta R_{\text{CLD}} + \Delta R_{\text{CO}_2} + \left( \frac{\partial R}{\partial T} \right) \Delta T \tag{2}
\end{align*}
\]

In Equation (2), the subscripts WV, CLD, CO2, and \( \alpha \) stand for water vapour, cloud, CO2 concentration, and surface albedo, respectively. \( \partial R / \partial T \) is the Planck feedback matrix whose \( j \)th column corresponds to the vertical profile of the radiative energy perturbation due to 1-K warming at the \( j \)th layer, and \( (\partial R / \partial T) \Delta T \) then represents the difference in vertical profile of the longwave radiation energy flux due to the changes in temperature itself. Substituting Equation (2) into (1) and then rearranging the terms we obtain:

\[
\Delta T = \left( \frac{\partial R}{\partial T} \right)^{-1} \left[ \Delta (S - R)_{\text{WV}} + \Delta (S - R)_{\text{CLD}} + \Delta (S - R)_{\text{CO}_2} \\
+ \Delta S_{\alpha} + \Delta Q_{\text{non-radiative}} \right] \tag{3}
\]

It should be noted that each \( \Delta (S - R)_X \), where \( X \) represents individual radiative processes such as water vapour, can be calculated as the difference between two calculations of a radiative transfer model: one with inputs from the mean state of 1979–1997 and the other with identical inputs except the field denoted by the subscripts, which is taken from the mean state of
1998–2016 over the TP. Also, the sum of non-radiative heating perturbations and heat storage terms \( \Delta Q_{\text{non-radiative}} \) is estimated according to the energy balance equation:

\[
\Delta Q_{\text{non-radiative}} = -\Delta (S - R)_{\text{total}} \tag{4}
\]

where \( \Delta (S - R)_{\text{total}} \) can be calculated as the difference between two calculations of the radiative transfer model, one with all input fields taken from the mean state of 1979–1997 and the other with all input fields from the mean state of 1998–2016. All the radiative heating calculations are made with the Fu–Liou radiative transfer model (Fu and Liou, 1992; 1993). Besides, at the atmospheric layers, \( Q_{\text{non-radiative}} \) represents the vertical profile of total energy perturbations due to the processes including convection, turbulence, and large-scale atmospheric motions \( (Q_{\text{atm, dyn}}) \). The surface component \( (Q_{\text{surf, dyn}}) \) of \( Q_{\text{non-radiative}} \) stands for the energy perturbations due to the changes in surface sensible flux \( (Q_{\text{SH}}) \), surface latent heat flux \( (Q_{\text{LH}}) \), and heat storage term \( (Q_{\text{heat, storage}}) \). Over ocean, \( \Delta Q_{\text{heat, storage}} \) includes oceanic heat transport and oceanic heat content. Over land, \( Q_{\text{non-radiative}} \) includes the energy loss or gain due to snow and ice melting/freezing and runoff. Then, we obtain:

\[
\Delta T = \begin{pmatrix} \frac{\partial R}{\partial T} \\ -1 \end{pmatrix} \left[ \Delta (S - R)_{\text{WV}} + \Delta (S - R)_{\text{CLD}} + \Delta (S - R)_{\text{CO2}} + \Delta S_r \right] + \Delta Q_{\text{SH}} + \Delta Q_{\text{LH}} + \Delta Q_{\text{heat, storage}} + \Delta Q_{\text{atm, dyn}} \tag{5}
\]

Based on Equation (5), the changes in SAT over the TP between the two climate states can be linearly decomposed into partial temperature changes (PTCs) due to the changes in radiative processes like water vapour, cloud, \( CO_2 \) concentration, surface albedo, and non-radiative processes such as surface sensible and latent heat fluxes, heat storage and atmospheric dynamics.

4 | RESULTS

4.1 | Attribution of annual mean warming and possible mechanism

Figure 2 shows the CFRAM-derived partial and total annual mean SAT anomalies between 1979–1997 and 1998–2016. Specifically, the sum of PTCs due to the changes in heat storage (Figure 2a), surface latent heat flux (Figure 2b), surface sensible heat flux (Figure 2c), surface albedo (Figure 2d), cloud (Figure 2e), atmospheric dynamic process (Figure 2f), \( CO_2 \) concentration (Figure 2g), and water vapour (Figure 2h) equals to the total SAT anomaly shown in Figure 2i. The spatial distribution of CFRAM-derived total SAT (Figure 2i, the shading) almost coincides the spatial pattern obtained from the ERA-Interim (Figure 2i, the contour), which adds more confidence to our decomposition analysis. The spatial distribution pattern of PTC due to the changes in heat storage leads to a warmer surface over the northern TP, especially over the northeast TP, where the warming rate is only smaller than that of surface sensible heat flux, but there is a strong cooling effect over the western TP (Figure 2a). The change in surface latent heat flux causes two cooling centres over the central and northeastern TP (Figure 2b). The changes in surface sensible heat flux, atmospheric dynamic process, and water vapour are the main contributors to the warming over the central and eastern TP (Figure 2c,f,h), while the changes in surface albedo and cloud lead to a cooling surface (Figure 2d,e). The PTC caused by changes in \( CO_2 \) concentration is uniformly positive at a magnitude about 0.1 K (Figure 2g), while changes in water vapour lead to a relatively warmer surface, especially in eastern and northern TP (Figure 2h). Besides, the changes in surface sensible heat flux, surface albedo, cloud, and atmospheric dynamic process lead to a contrastive PTC between the platform and higher elevation regions of the TP. To be specific, the changes in surface sensible heat flux and atmospheric dynamic process contribute to a warming surface over the central and eastern TP but a cooling surface over the western part and southern flank of the TP. On the contrary, the PTCs due to changes in surface latent heat flux, surface albedo, and cloud are negative over the central and eastern part but positive over the western part. The PTC due to the changes in surface sensible heat flux that warms the surface is largely compensated by surface albedo cooling over the central and eastern TP, while over the western TP, things are just opposite. Overall, the combined effect of these processes leads to consistent warming over the entire TP.

For annual mean, surface albedo shows an increase (Figure 5a) and surface sensible heat flux reveals a reduction (Figure 5d) over the central and eastern TP between 1979–1997 and 1998–2016, but over the western area and southern slope, surface albedo displays a more remarkable drop and surface sensible heat flux manifest a comparable growth. These changes in surface albedo have been illustrated by Wang et al. (2018) based on the data obtained from the National Snow and Ice Data Center of 1979–2006, who found that the linear trends of snow cover experienced a remarkable decrease in the western and southern TP, but a significant increase in the central and eastern parts. Besides, Duan and Wu (2008) showed that, based on the station data, surface sensible heat flux experienced decreasing trends in all seasons over the
central and eastern TP during 1980–2003 due to the sharply reduced wind speed, which was confirmed by other studies (Yang et al., 2010; Duan et al., 2011; Zhu et al., 2017). Zhu et al. (2017) also presented a recovery of surface sensible heat flux since 1998 due to the increased difference in surface and air temperatures induced by increase in total cloud cover. The decreased surface albedo makes more solar radiation reaching to the surface, while the decreased surface sensible heat flux makes less energy transfer from the surface to the air. The combined effects would allow more energy to remain at the near-surface layer to warm the surface. Moreover, the column-integrated water vapour change shows a remarkable convergence over the central TP and a divergence over the southern flank (Figure 6a). Station data also showed a positive trend of the specific humidity at the surface layer for a shorter analysis period (1984–2006; Yang et al., 2014). Meanwhile, an anomalous divergence circulation appears at the upper level over the southern flank but a weak uplifting movement over the central and eastern TP (Figure 6d). The upward motion, together with the increased water vapour in the atmosphere, is favourable for more cloud cover, leading to a cooler surface due to the reduction in surface net solar radiation (Figure 2e).

4.2 Attribution of summertime warming and possible mechanism

The contributions of individual processes to SAT changes in JJA mean between the two stages are demonstrated in Figure 3. The change in heat storage shows an obvious negative effect over the western TP but a positive effect over the central and eastern TP (Figure 3a). The contributions caused by the changes in surface latent heat flux (Figure 3b) and surface sensible heat flux (Figure 3c) can be well compensated by each other. Compared with the weak cooling effect caused by the change in surface albedo in annual mean (Figure 2d), the reduction in
surface albedo in JJA mean has a stronger warming effect on the central and eastern TP SAT (Figure 3d). The spatial distribution pattern of PTCs due to the changes in cloud, atmospheric dynamic process, CO₂ concentration, and water vapour (Figure 3e–h) are quite similar with that in annual mean (Figure 2e–h), but generally with a larger magnitude. Specifically, the change in cloud acts to cool the surface larger than 2 K, while the changes in surface albedo and surface sensible heat flux tend to warm the surface about 1–1.5 K, and the contribution is about 0.5–1 K to total surface warming due to atmospheric dynamic process and water vapour feedback over the central and eastern TP. Figure 3i illustrates that there is a stronger warming over the northern region compared with the rest of the TP, which is the combined effect of surface sensible heat flux, surface albedo, and water vapour changes (Figure 3c,d, and h).

Figure 5b shows the significant increase in cloud between the two climate stages by given the column-integrated cloud ice and liquid water content over the central and eastern TP. The remarkable increased cloud may be related to the strong water vapour convergence (Figure 6b) and the anomalous ascending motion (Figure 6e) over this area. The increase in cloud has different influences on the SAT. On the one hand, more clouds can reflect more solar radiation, resulting in less shortwave radiation reaching to the surface. On the other hand, clouds can trap more longwave radiation between the clouds and the surface, causing a warmer surface. In general, the shortwave effect is dominant at the surface but the longwave effect is dominant in the overlying air. So, more cloud cover is favourable for the surface cooling but atmospheric warming. Given that the TP surface is the heating source in summer, cloud-induced warmer air and colder surface result in a smaller difference between the surface and air temperatures, leading to a decreased surface sensible heat flux (Figure 5e). In general, even though more cloud cover induced by the enhanced water vapour content and ascending motion would lead to a cooling surface, it can be well compensated by the warming due to reduced surface sensible heat flux.

4.3 | Attribution of wintertime warming and possible mechanism

The spatial distributions of DJF mean PTCs are shown in Figure 4. Generally, the most important process for
warming over the entire TP in winter is the changes in heat storage, surface latent heat flux, surface sensible heat flux, and surface albedo (Figure 4a–d), while the contributions of changes in cloud (Figure 4e), atmospheric dynamic process (Figure 4f), and water vapour (Figure 4h) are negligible. The changes in heat storage and surface sensible heat flux act to warm the surface air at the value of more than 1 K (Figure 4a,c), while change in surface albedo tends to cool the surface at the rate more than $-1$ K (Figure 4d) in most of the central and eastern TP. In the western region and the southern flank of the TP, the changes in surface latent heat flux and surface albedo are the main contributors for the increasing SAT. Contributions due to the changes in surface sensible heat flux and surface albedo can largely be compensated by each other over the entire TP. Figure 4i shows that, in winter, the warming in the northeastern area and southern flank region is most obvious compared with the rest region of the TP, which is consistent with the spatial distribution of tendency in Figure 1i.

Figure 5c shows that surface albedo increases remarkably in winter over the central and eastern TP, which can also be seen in Wang et al. (2018). Moreover, A quite new study (Bao and You, 2019) pointed out that the variation and meridional shifts of westerly jet significantly influence the snow accumulation over the TP in the late winter by promoting ascending motion and water vapour convergence from the southwest. Here the increased westerly winds (Figure 6f) may contribute to more snow cover (Figure 5c), even though the increase in water vapour flux is not obvious (Figure 6c). The surface with more snow/ice cover reflects more solar radiation, leading to a reduction surface net solar radiation and a colder surface (Ghatak et al., 2014; You et al., 2016). However, the decadal reduction of surface sensible heat flux (Figure 5f) remains more energy at the near-surface layer, indicating a warmer surface, which weakens the cooling due to the change in surface albedo.

### 4.4 Areal mean results over the TP

To obtain the quantitative contributions of individual processes to the long-term changes in SAT over the entire TP, we summarize the results in Figures 2–4 by calculating the area-mean total and partial SAT anomalies.
As shown in Figure 7a, surface albedo has a strong positive contribution (0.32 K) in JJA mean, but a negative contribution (−0.19 K) in DJF mean. Water vapour feedback is important in regulating annual and JJA mean SAT changes over the TP, with a contribution at 0.28 and 0.46 K, respectively. However, the change in
cloud has a strong cooling effect on the surface at the value of \(-1\) K in JJA mean, while in annual and DJF means, the contribution of cloud change is close to zero since the cooling on the central and eastern parts can be well compensated by the warming over the western and southern regions. This compensation is also the reason why the contributions due to the changes in atmospheric dynamic process are very small, especially in annual and JJA means. The greenhouse effect, which can be represented by the increase in CO\(_2\) concentration, adds almost 0.2 K warming over the entire TP in annual mean SAT change, as well as in JJA mean and DJF mean. The change in heat storage has strong warming effect (about 0.6 K) in wintertime. The reduction in surface sensible heat flux favours SAT increase in all three situations. Considering the significant contrastive phenomenon over the plateau region and to the west, the areal average east of 80\(^\circ\)E is also examined (Figure 7b).

\[
\begin{align*}
\text{FIGURE 7} & \quad \text{(a) Areal mean of entire TP} \\
\text{FIGURE 7} & \quad \text{(b) Areal mean of east of 80\(^\circ\)E}
\end{align*}
\]

\[
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\]

This study attributes the long-term warming over the TP in the past four decades. By applying the process-based decomposition method CFRAM, we quantitatively analyse the relative contributions of the external forcing and internal climate processes to the warming over the TP in annual mean, JJA mean, and DJF mean. It is found that the warming in annual mean (0.62 K) is mainly contributed by the reduction in surface sensible heat flux (0.29 K) and increase in water vapour (0.29 K), which allow more energy to remain at the near-surface layer and lead to an increase in SAT. In JJA mean, although a significant increase in cloud causes surface cooling (\(-1.0\) K), which is compensated by surface warming due to reduced surface sensible heat flux (0.48 K) and a more remarkable increasing trend in water vapour (0.46 K). The surface albedo feedback further amplifies surface warming about 0.32 K by increasing net surface shortwave radiation. The combined effect of all the processes makes a warming surface over the TP in summer...
(0.47 K). In winter, the contributions of the changes in water vapour and cloud to the long-term trend of SAT are negligible. The change in surface albedo causes a surface cooling about −0.19 K, which is compensated by surface warming related to the decreased surface sensible heat flux (0.26 K). Especially, the change in heat storage term accounts for 0.59 K surface warming. The combination of all processes makes a surface warming of 0.58 K in winter.

By using the CFRAM, we have quantitively analysed the possible mechanisms for the pronounced warming. It should be noted that although the ERA-Interim is one of the best reanalysis datasets to describe the climate change over the TP, it has not reached consensus with the other datasets on the long-term changes in certain variables, especially the cloud (Duan and Xiao, 2015), surface albedo, and surface heat fluxes (Su et al., 2017). According to Duan and Xiao (2015), the station data showed that the total cloud cover displayed a remarkable decreasing trend at both daytime and nighttime, but the low-level cloud experienced a weak increasing trend over the central and eastern TP. Here we have just focused on the combined radiative effect of the total cloud. Based on the MERRA, surface albedo exhibits a negative trend over the entire TP, which is opposite from the trend in the ERA-Interim. Moreover, due to the limitation of data, we cannot isolate the contributions of the land use and aerosol to the warming over the TP (Cui et al., 2006; Jin et al., 2010; Jiang et al., 2017), which are lumped up together with terms representing the dynamic processes in this study.

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