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Recent recovery of the boreal spring sensible heating over the

Tibetan Plateau will continue in CMIP6 future projections

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Abstract

The spring sensible heating (SH) over the Tibetan Plateau (TP) serves as a huge 'air pump', significantly influencing the Asian summer monsoon, has experienced a decreasing trend. However, it remains unclear whether this decline will continue. Therefore, we here examine the long-term trends of spring SH over the central and eastern TP (CETP) based on a meteorological station-based calculated SH dataset, and CMIP6 multi-model simulations. These two sources confirmed the previous finding that the SH peaks in May. Further, we find that the declining SH was replaced by a fast recovery after approximate 2000 in the station-based SH. This is to some extent verified by the historical simulations of CMIP6 models. Importantly, CMIP6 future projections suggest that this increasing trend will continue, and get stronger with higher radiative forcing from SSP126 to SSP585. Mechanism analysis indicates that the previous decreasing trend in SH was mainly caused by the decline of 10 m wind speed, while the recent and future increasing trend results from the rising ground-air temperature difference. We suggest that this increasing trend of spring SH over the CETP may serve as an alternative driver for the enhancement of the East Asian summer monsoon in the future.

1. Introduction

In boreal spring and summer seasons, the Tibetan Plateau (TP) directly heats the middle atmosphere of the troposphere, serving as a huge 'air pump' (Flohn 1957, Ye *et al* 1957, Li and Yanai 1996). Further studies (Liu *et al* 2002, Wu *et al* 2015, Wu *et al* 2018) suggested that this 'pump' was driven by the surface sensible heating (SH), which accounted for the abrupt change of the Asian atmospheric circulation from winter to summer, contributed to maintain the Asian summer monsoon (He *et al* 1987, Ueda and Yasunari 1998, Hsu and Liu 2003, Wu *et al* 2012, Duan *et al* 2018a, Yao *et al* 2019), and even regulated the climate change in Asia (Zhao and Chen 2001, Duan *et al* 2003, Wang *et al* 2014). Several

demonstrated that the phase characteristics of the Asian monsoon onset were consistent with the variations of the SH over the TP, reflecting the critical role of the TP SH on the Asian monsoon onset. Therefore, insights into the variations of the TP SH are crucial in understanding the changes of the Asian summer monsoon. In recent years, the SH over the TP has been proved

studies (Wu and Zhang 1998a, 1998b, 1999) have

In recent years, the SH over the 1P has been proved to be undergoing a decreasing trend, especially in spring season (Duan and Wu 2008, Wang *et al* 2012, Wang and Ma 2018). Duan and Wu (2008) suggested that this weakening trend in SH was induced mainly by the steadily decreased surface wind speed, which was linked to climate warming (Duan and Wu 2009). Moreover, the trends in SH over the TP also shouldered the important





Figure 1. Geographical distributions of 7/ meteorological stations over the CE1P. The direct comparisons between the eddy covariance system observed (denoted by 'TIPEX-III') and meteorological station-based calculated SH (denoted by 'Duan') at Amdo, Naqu, and Bange stations (red stars) are also presented here (units: W m⁻²). The boxed area indicates the domain (27–39 °N, 85–105 °E) used for area averaging of the CMIP6 model simulations in the following text.

responsibility for the changes in precipitation over Asia (Liu and Duan 2017). The decadal weakening in spring SH over the TP can lead to the inhibition of the SH pump; consequently, the following summer precipitation on the southern and eastern slopes of the TP was decreased, but those in northeast India, the Bay of Bengal (Duan *et al* 2011), south and north China (Duan *et al* 2013) were increased.

It is well established that the TP undergoes a significant temperature increase under global warming (Liu and Chen 2000, Zhou and Zhang 2005, Chen et al 2015, Ma et al 2017), with an accelerating warming trend especially since 1998 (Duan and Xiao 2015, You et al 2016, Yao et al 2019). In contrast to the increasing temperature, the SH over the TP has been decreased. However, it remains unclear whether this weakening trend of TP SH will continue associated with global warming, which is critical to understand the current and future changes in the Asian summer monsoon. Therefore, in this study, we comprehensively examine the changes of the TP SH trend based on a meteorological station-based calculated dataset, and the phase 6 of the Coupled Model Intercomparsion Project (CMIP6) historical simulations and future predictions for the IPCC sixth Assessment Report (AR6). Further, the direct mechanisms behind these changes of the TP SH trend are investigated.

2. Datasets, model simulations, and methodology

2.1. Datasets used

The SH, air temperature (Ta), ground surface temperature (Ts) and 10 m wind speed (V_{10}) datasets used here are provided by Duan *et al* (2018b). They have been publicly released, and can be downloaded and used freely (http://staff.lasg.ac.cn/amduan/index/ article/index/arid/11.html). This set of meteorological observations originated from the China Meteorological Administration (CMA). In this study, we employed the above data at 77 stations (figure 1) over the central and eastern TP (CETP) from 1979 to 2014.

The SH fluxes at several sites directly observed by the eddy covariance system in the Third TP Atmospheric Scientific Experiment (TIPEX-III, Zhao *et al* 2018) were also adopted in this study. These observations have been shared at 'the special site on the basic data of atmospheric science experiments over the TP' (http://123.56.215.19/tipex).

Global Precipitation Climatology Center (GPCC, https://dwd.de/EN/ourservices/gpcc/gpcc.html) gridded monthly precipitation dataset from 1979 to 2014 was used here, which was calculated from global station data (Schneider *et al* 2011, https://esrl.noaa.gov/psd/data/ gridded/data.gpcc.html).

2.2. Model simulations

We adopted CMIP6 multi-model simulations (https:// esgf-node.llnl.gov/projects/cmip6/) in this study. Currently, 12 models in CMIP6 provide the historical simulations from 1850 to 2014, while only 6 of them (BCC-CSM2-MR, CNRM-CM6-1, CNRM-ESM2-1, IPSL-CM6A-LR, MRI-ESM2.0, and UKESM1.0-LL) provide future climate projections (O'Neill *et al* 2016) until 2100 under different emission scenarios. These alternative scenarios are driven by a new set of emissions and land use scenarios (Riahi *et al* 2016) based on new future pathways of societal development, the Shared Socioeconomic Pathways (SSPs), and related to the



Table 1. List of models used in this study, including the model names, countries, numbers of ensembles, resolutions, and data references. All of the models include historical simulations, and six of them (in bold) provide the future climate projections under the alternative scenarios.

No.	Model	Country	No. of ensembles	Resolutions	Data references
1	BCC-CSM2-MR	China	3	160 × 320	Beijing Climate Center (BCC) 2017a
2	BCC-ESM1	China	3	64 imes 128	Beijing Climate Center (BCC) 2017b
3	CESM2	USA	10	192×288	Danabasoglu 2019b
4	CESM2-WACCM	USA	3	192×288	Danabasoglu 2019a
5	CNRM-CM6-1	France	10	128×256	Voldoire et al 2019
6	CNRM-ESM2-1	France	5	128×256	Seferian 2018
7	GISS-E2.1G	USA	10	90×144	NASA Goddard Institute for Space Studies (NASA/
					GISS) 2018
8	GISS-E2.1H	USA	10	90×144	NASA Goddard Institute for Space Studies (NASA/
					GISS) 2017
9	IPSL-CM6A-LR	France	31	143×144	Boucher et al 2018
10	MIROC6	Japan	10	128×256	Tatebe and Watanabe 2018
11	MRI-ESM2.0	Japan	5	160×320	Yukimoto et al 2019
12	UKESM1.0-LL	UK	6	144×192	Met Office Hadley Centre 2017

Representative Concentration Pathways (RCPs, van Vuuren *et al* 2011). The SSPs and the associated scenarios presented here are SSP126, SSP245, SSP370 and SSP585, with low, medium, a medium to high, and high radiative forcing by the end of century, respectively (Meehl *et al* 2014, O'Neill *et al* 2014, van Vuuren *et al* 2014). All of the models and their number of ensembles are listed in table 1. Given that their different horizontal resolutions (table 1), we interpolated their simulated variables consistently into $1.5^{\circ} \times 1.5^{\circ}$ based on the first-order conservative remapping scheme (Jones 1999) by Climate Data Operators (CDO).

2.3. Methodology

According to previous studies (Monin and Obukhov 1954, Duan and Wu 2008, Wang *et al* 2012, Duan *et al* 2018b), SH can be calculated by the following equation:

$$SH = \rho C_p C_H V_{10} (T_s - T_a)$$
(1)

where ρ is the air density, C_p is the specific heat of dry air at constant pressure, V_{10} is the mean wind speed at 10 meters above the surface, and Ts-Ta is the groundair temperature difference. C_H , named bulk transfer coefficient for heat, is a complicated variable with strong diurnal variation, which depends on the atmospheric stability, dynamical and thermal roughness (Yang *et al* 2010) and wind speed (Chen *et al* 1985, Wang *et al* 2016). Duan *et al* (2018b) calculated the SH data used in this study by taking into account that C_H is changing according to the scheme proposed by Yang *et al* (2010). More details can be referred to Yang *et al* (2010) and Duan *et al* (2018b).

In order to verify the reliability of this stationbased calculated SH dataset from Duan *et al* (2018b), we picked up some SH fluxes directly observed by the eddy covariance system in TIPEX-III at the same locations and made the comparison (figure 1). We can see that the SH from Duan *et al* (2018b) basically matches well with the directly observed SH for all three stations. These two sets of SH data have comparable amplitudes and consistent variations at Amdo and Naqu stations, with the respective correlation coefficients of 0.62 and 0.82, and root mean square errors of 15.79 and 11.58 W m⁻². At Bange station, although they somewhat differ in amplitudes, they have the almost same month-to-month variations, with a higher correlation coefficient of 0.95. Therefore, the station-based calculated SH from Duan *et al* (2018b) used in this study can be trustworthy, and hereafter it was referred to as the observations.

For CMIP6 simulations, the median of the multimodel simulations was calculated, which has the advantage of avoiding influence by the extremely large or small values. And the piecewise linear regression is applied here as a statistical technique to detect the turning point in the trend changes (Toms and Lesperance 2003, Ryan and Porth 2007).

3. Results

3.1. Spring SH over the CETP

Many literatures have illustrated that spring SH over the TP played an important role in the Asian summer monsoon (Wu and Zhang 1998a, Zhao and Chen 2001, Wu *et al* 2018, Yao *et al* 2019). More specifically, we first present the annual cycle of SH over the CETP and the influence of spring SH over the CETP on the summer precipitation patterns over East Asia in figure 2.

In figure 2(a), the series are obtained by site averaging for 77 meteorological stations (blue markers in figure 1) and area averaging for gridded products (within the blue box in figure 1), respectively. Though differences in the amplitudes of SH over the CETP exist between two different datasets which both have uncertainties, the annual variations are generally consistent. The station-based SH shows that the SH peaks in spring with the maximum value in May, which has a close relationship with the changes in circulation over the surrounding areas (Li 1987, Wu *et al* 2018).





CETP during 1979–2014 (pink line). The gray and blue bars indicate the precipitation in observations and CMIP6 simulations, respectively. The pink shaded areas and the error-bars represent the areas between the 25% and 75% percentiles. The inserted bar chart on the top right describes the month of the maximum SH value in each CMIP6 model. (b) Regression coefficients of the summer precipitation against the spring SH over the CETP (units: mm mon⁻¹/W m⁻²), and dotted areas exceed the 90% confidence level.

Further, SH gradually decreases after the outbreak of the TP summer monsoon owing to more precipitation, which is consistent with previous findings that the surface and air temperature difference reaches its peak in May before the arrival of the monsoon (Liao *et al* 2019). For the CMIP6 model outputs, the median SH among the 12 models is obviously underestimated from January to May, making the SH in summer seem larger than that in spring. This may be caused by the wet bias in the simulated precipitation during the premonsoon season (figure 2(a)). However, the maximum SH in median still occurs in May, in accordance with the observations. More specifically, five out of the 12 models exhibit the maximum SH in May, while the other seven models show maximum SH in June.

Figure 2(b) clearly shows that spring SH obviously affects the summer precipitation patterns over the East Asia region. Considering the weakening trend in SH over the CETP examined in previous studies (Duan and Wu 2008, Wang *et al* 2012), we know that the summer precipitation in southern China and Indo-China Peninsula actually underwent a rising trend, while the opposite condition mainly occurred in central China during the past decades, which was consistent with the results of Duan *et al* (2013). Therefore, in order to better understand and predict the summer precipitation changes in East Asia, it is necessary for us to continuously monitor the long-term variations of spring SH over the CETP.

3.2. Recent recovery of spring SH over the CETP

Considering the dominance and great importance of spring SH over the CETP, we show the long-term trend of the SH in figure 3. Based on the meteorological station observations (figure 3(a)), we clearly find that the SH was significantly weakened at a rate of -1.93 W m^{-2} per decade (p < 0.01) during the first two decades, as suggested in previous studies (Duan and Wu 2008, Yang *et al* 2010). However, a recovery occurs with a significant increase at a rate of 1.87 W m^{-2} per decade (p < 0.05) in recent decades. We identify this turning point in 2000 by the piecewise linear regression, which is consistent with the results of Zhu *et al* (2017).

In terms of the historical simulations in CMIP6 (figure 3(b)), the medians of SH among the 12 models and six models with future scenarios can both reproduce the features in figure 3(a), with a significant decreasing trend in the first two decades and an increasing trend in recent decades. And the turning point is detected around 1998, which is very close to 2000 shown by the observations. Collectively, although the absolute amplitude of SH differs among different data products, their variations are broadly consistent, showing a reversal from the significant





decreasing trend to a remarkable increasing trend sur

3.3. Continued increasing trends in CMIP6 future

under global warming.

projections

An interesting question arises as to whether this increasing trend in SH will continue in the future? Here we employed the future projections simulated by the six models in CMIP6 to address this question. Clearly, figure 3(b) suggests that the SH will continue to show an obvious increasing trend of 0.81 W m⁻² per decade (p < 0.01) from 2015 to 2100 in the CMIP6 future projection under the scenario SSP585.

As we know, global warming is projected to become stronger from 2015 to 2100, and more significant warming over the TP regions are expected under more radiative forcing from the scenario SSP126 to SSP585 (figure S1 is available online at stacks.iop.org/ERL/14/ 124066/mmedia). Under these conditions, the CMIP6 models consistently suggest that the rising trends in SH will get stronger along with the stronger global warming from SSP126 to SSP585 (figure 3(c)). These future projections under different scenarios indicate that the increasing trend of SH over the CETP will be a longterm behavior rather than a decadal change. Moreover, this increasing trend will enhance the East Asian summer monsoon, favoring more precipitation over central China and less over south China (figure 2(b)), which is an opposite precipitation pattern of the first two decades (from 1979 to around 2000).

3.4. Causes of the changes in SH over the CETP

For brevity, the formula of SH (equation (1)) can be expressed as:

$$SH = CV_{10}(T_s - T_a) \tag{2}$$

where $C = \rho C_p C_H$ is called here the C coefficient, which represents a synthesis of several parameters, especially the transfer coefficient for heat (C_H). Therefore, according to the equation (2), we can deduce the trend of spring SH over the CETP (dSH/dt) in terms of the trends of C coefficient (dC/dt), 10 m wind speed (dV_{10}/dt), and ground-air temperature difference ($d(T_s - T_a)/dt$).

Firstly, we examine the evolution of the C coefficient in the station observations during 1979–2014, and find that it shows an increasing trend before 2000 and a decreasing trend after that (figure S2(a)). In addition, the changes of the V₁₀ and Ts-Ta in the station observations during recent three decades are illustrated in figure 4(a). It clearly reveals that V₁₀ dropped sharply at a rate of -0.33 m s⁻¹ decade⁻¹ (p < 0.01) in the first two decades, but almost levelled off (-0.02 m s⁻¹ decade⁻¹) in





the last decade, with a turning point at 2001. In contrast, the Ts-Ta demonstrates a nearly opposite change against the V_{10} , with a levelling off before 1999 and thereafter a significant increasing trend at a rate of 0.53 °C decade⁻¹ (p < 0.01). Comparing the variations of SH (figure 3(a)) and these three terms (figures 4(a) and S2(a)), we can easily conclude that the distinct decreasing trend in SH before 2000 can be accounted for by the obvious decline in V_{10} , which confirms the result revealed by previous studies (Duan and Wu 2008, 2009). However, after 2000, the Ts-Ta shows a prominent rise (figure 4(a)), acting as the dominant driver for the increasing trend in SH. And the increase in Ts-Ta after 2000 is caused by the fact that the warming in Ts is more rapid than the Ta under more radiative forcing along with global warming. Therefore, the transition in mechanism from the decline in V₁₀ to the rapid rise in Ts–Ta can easily account for the reversal of the changes in spring SH over the CETP before and after 2000.

In the CMIP6 historical simulations, the simulated V_{10} also indicates a decreasing trend, while the Ts-Ta shows an increasing trend (figure 4(b)). These changes are generally in accord with the meteorological station observations, though there are no obvious turning points. Given that an important cause of the decreased wind is probably changed by surface roughness (Zhang *et al* 2019), which is not considered in the model, this result is surprising. And Lin *et al* (2013) have pointed out that circulation change instead of roughness change was the major cause of wind decline

over the TP. Additionally, the variations of C coefficient in the CMIP6 models also exhibit a weak decline in recent three decades (figure S2(b)). In comparison, we can also infer that the decreasing trend of the simulated SH before 1998 mainly results from the decreasing trends of the V₁₀ and C coefficient, but the increasing trend is explained by the increasing trend of Ts–Ta (figure 4(b)). Moreover, though only six CMIP6 models' simulations are shown in figure 4(b), the 12 CMIP6 models' historical simulations suggest the identical variations in V₁₀ and Ts–Ta. Therefore, in a word, the mechanisms for the reversal in SH revealed by the CMIP6 model simulations coincide with the station observations.

After 2014, the simulated V₁₀ and Ts–Ta continue to significantly decrease and increase under SSP585 scenario, respectively (figure 4(b)). The slopes of their trends over the entire period of 1979–2100 are -0.03 m s^{-1} decade⁻¹ and 0.08 K decade⁻¹ (p < 0.01), respectively. Therefore, the continuous increase of SH under SSP585 can still be attributed to the rise of the Ts–Ta, which is actually forced by the more net downward shortwave radiation (figure 4(b)). And the more net downward shortwave radiation is a result that is complicatedly influenced by the changes of aerosols, clouds, and surface albedos.

Further, we showed the trends of V_{10} and Ts–Ta from 2015 to 2100 under four different scenarios in figures 4(c) and (d). We can see that the variations in V_{10} under all the scenarios demonstrate the decreasing trends, while variations in Ts—Ta reflect the increasing trends. More importantly, the strengths of their trends get enhanced from SSP126 to SSP585. Hence, the increasing trends of the SH under different scenarios (figure 3(c)), especially from medium to high radiative forcing by the end of century (SSP245, SSP370, and SSP585), are consistently dominated by the rising Ts—Ta (figure 4(d)).

4. Conclusion and discussion

In this study, we analyzed the variation of boreal spring SH over the CETP and its possible mechanisms based on the meteorological station-based calculated data, and CMIP6 historical simulations and future projections. We find that the station-based calculated SH peaks in boreal spring with the maximum value in May. The CMIP6 models simulated SH median is somewhat underestimated, but the maximum value also occurs in May.

We further suggest that there is a recovery of the declining SH during 1979 and 2014 with the turning point at approximate 2000 in both datasets. The CMIP6 model future projections under the four scenarios demonstrate that this increasing trend will continue, and the trend will get stronger from SSP126 to SSP585. The station observations and the CMIP6 simulations both indicate that the decline of the wind speed at 10 m plays an important role in the decreasing trend of SH, while the rising ground-air temperature difference contributes to the increasing trend in the recent decade and even in the future warming world.

It should be kept in mind that accurate calculation and assessment of SH over the TP remains a great challenge. As we know, CMA meteorological stations are distributed irregularly with more stations located at low elevation than high elevations. And the Ts observed at the meteorological station basically reflects the ground temperature of the bare soil, which is poorly represented for the vegetated regions over the CETP (Yang et al 2010, Liao et al 2019). Additionally, different choices of C_H values also have the important effect on the strength of calculated SH. Many previous literatures took C_H as a constant to calculate the SH over the TP (Li and Yanai 1996, Li et al 1996, Duan and Wu 2008). However, C_H is actually variable, which depends on the atmospheric stability, dynamical and thermal roughness (Yang et al 2010) and wind speed (Chen et al 1985, Wang et al 2016). Wang et al (2016) suggested that C_H was 2.4×10^{-3} over the western part of the TP, 2.2×10^{-3} - 3.8×10^{-3} in the middle of the TP, and 3.4 \times 10⁻³–6.0 \times 10⁻³ in the eastern and southern side of the TP.

In terms of the CMIP6 simulations, it is not surprising that large inter-model spread exists, owing to their different model structures, physical parameterizations, horizontal and vertical resolutions, and so forth. Additionally, it is worth mentioning that greening is occurring over the TP (Shen *et al* 2015).



But these vegetation changes are not considered in the climate system models, and their effects on the trend of SH over the TP remain unclear. Furthermore, the validity of SH products over the TP in the reanalysis database remain debatable (Zhu *et al* 2012). The changes in SH in two newly released reanalysis products, ERA5 (Copernicus Climate Change Service (C3S) 2017) and JRA-55 (Kobayashi *et al* 2015), are also investigated (figure S3). Both of them show that SH variations over the CETP reflect an increasing trend, which disagrees with the observations, and CMIP6 simulations. Therefore, we should pay continuous attention to how to constrain SH strength over the TP and its trend in the future, because of its vital influences on the Asian summer monsoon.

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Data availability statements

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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