Numerical simulation of Tibetan Plateau heating anomaly influence on westerly jet in spring

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The intensity and location of the westerly jet (WJ) in the mid-latitudes of the Northern Hemisphere has been closely related to weather and climate changes in Eurasia. In this paper, we analyzed the seasonal and inter-annual variations in the strength and position of the WJ, using the NCEP-DOE reanalysis II (NCEP) data. Spring is the period that the Tibetan Plateau (TP) transits from a cold to a warm condition, and at the same time the WJ weakens and moves northward. Correlation analysis showed that the westerly wind over the north (south) of the TP has a significant positive (negative) correlation with the TP surface temperature in spring during 1979–2013 (99.9% significance level). In spring, surface sensible heat release dominates thermal conditions of the TP, while the contribution of latent heat becomes important in summer. We conducted three numerical experiments to determine the effects of the TP surface sensible heat flux anomaly in spring on the intensity and location of the WJ using the community atmosphere model 5 (CAM5). Experimental results indicated that changing the TP surface sensible heat flux can produce changes in the TP heat sources. When the TP sensible heat flux is switched off (TPSHL), the TP surface has a lower temperature and no longer heats up the middle and upper troposphere, and leads to a maximum cooling of -1.2°C at 200 hPa over the TP with a significance level of 95%. In spring, the abnormal low temperature over the TP caused the WJ to strengthen and move southward compared with the control experiment (CTL), and this results in a westerly (easterly) wind anomaly over the south (north) of the TP with a significance level of 95%. We obtained the opposite conclusion from the comparison of the enhanced TP surface sensible heat flux (TPSHH) with CTL test results. In TPSHL, the WJ is still steadily located near 30°N in May as well as April, i.e., the first northward jump occurs from May to June which is a delay of one month compared to CTL. While in TPSHH, the WJ moves northward approximately 10° of latitude from April to May which is 5° larger than CTL. Thus, the strength and north–south migration of WJ is closely linked to the TP heat source, especially during the spring–summer seasonal transition period, while TP surface temperature increased significantly and showed early signal compared to the surroundings with the global warming. Therefore, considerable attention must be paid to the location and strength of the WJ with respect to the sensible heat flux over the TP.

1. Introduction

The northern hemisphere mid-latitude westerly jet (WJ) is an important component of the global

climate system and is a key link between high latitudes, the North Atlantic and the East Asian monsoon regions. The WJ intensity and position changes play a critical role in the climate change

Keywords. Tibetan Plateau; heat source anomaly; westerly jet; numerical simulation.

of Eurasia (Porter and An 1995; Zhang *et al.* 1997; Sun *et al.* 2012; Li *et al.* 2013). Changes in the WJ are expected to lead to a greater adjustment of the entire Asian monsoon pattern (Mölg *et al.* 2013; Nagashima *et al.* 2013; Li *et al.* 2014). The meridional location of the WJ alternates between the north and south of the Tibetan Plateau (TP) due to the thermodynamics of the TP. Under the influence of solar radiation, the WJ is strong and flows along the south of the TP during the winter. The WJ weakens and moves to the north of the TP in summer (Schiemann *et al.* 2008). Numerous studies suggest that the seasonal transition of the atmospheric circulation, including the onset/retreat of the Asian monsoon and the advance of the rainfall belt over East Asia, has close relationships with the changes in the position and strength of the WJ (Chen 1957; Zhang *et al.* 2006; Sampe and Xie 2010). Chen (1957) showed that the onset of Meiyu and the southwest monsoon coincides with the timing of the WJ shift from south to north of the TP. Yang *et al.* (2002) argued that the WJ seems to be more strongly linked to the climate changes of the Asia and Pacific regions than ENSO. Observational evidences suggest that southward shifts in the summer cooling of the upper level of the WJ over East Asia is accompanied by a weakening in the East Asian summer monsoon. This results in a tendency towards increased droughts in northern China and flooding along the Yangtze River valley (Yu *et al.* 2004). According to the principle of thermal wind balance, the WJ changes are mainly modulated by the temperature gradient caused by the TP surface heating (Krishnamurti 1961). Therefore, the relationship between the WJ and TP heat source is very important and worthy of further analysis using observation data and numerical models.

The TP is known as the ‘roof of the world’ and is subjected to strong interactions among the atmosphere, hydrosphere, cryosphere and biosphere (Yang *et al.* 2014). The TP acts as a heat source (sink) and thereby has an extremely important influence on the Eurasian climate pattern and even the global atmospheric circulation (Flohn 1957; Ye *et al.* 1957; Ding 1992; Wu *et al.* 2007). The TP heat source anomaly and its effects have gathered widespread attention in the scientific community. Regarding TP heat source effects, scientists have done a lot of researches, and achieved fruitful results. Generally speaking, the TP is called ‘the sensitive area’ and ‘the start-up region’ in Asia (Feng *et al.* 1998; Yao *et al.* 2012) and is characterized as ‘the driving force’ and ‘the amplifier’ for global climate change caused by the TP thermodynamics (Liu and Chen 2000; Pepin *et al.* 2015). The TP exerts a huge thermal forcing on the mid-troposphere over the middle-latitude of the Northern Hemisphere during spring and summer

(Flohn 1957; Ye and Gao 1979). The TP thermal forcing effectively enhances the Asian summer monsoon and modulates its variability (Yanai *et al.* 1992; Wu and Zhang 1998), and it also triggers vigorous deep convections over the TP (Yang *et al.* 2014). The previous studies have mainly investigated how the TP surface heat anomaly influences the onset and development of the Asian summer monsoon using observational data from meteorological stations, reanalysis and remote sensing (Luo and Yanai 1984; Wu and Zhang 1998; Liu and Chen 2000; Duan *et al.* 2005; Zhao *et al.* 2007; Zhou *et al.* 2009; Liu and Wang 2011) and model simulations (Chen *et al.* 2003; Li and Liu 2014). Wu and Zhang (1998) used data from meteorological stations to confirm that the whole process of the onset of the Asian monsoon is composed of three consequential stages. The first is the monsoon onset over the eastern coast of the Bay of Bengal (BOB) in early May. This is followed by the onset of the East Asian monsoon over the South China Sea (SCS) by 20 May, and then the onset of the south Asian monsoon over India by 10 June. They argue that the onset process of Asian monsoon is directly linked to the thermal and mechanical forcing of the TP. The role of TP on the withdrawal of the Asian summer monsoon has also received significant attention. The sudden withdrawal of the Indian summer monsoon coincides with abrupt temperature drops over the Eurasian continent and southward shifts of the WJ around the TP (Matsumoto 1992).

In short, the thermal influences of TP upon atmospheric circulation are of great importance. Because spring is the period in which the TP transits a cold to a warm condition, surface sensible heat release dominates thermal conditions of the TP in spring, while the contribution of latent heat becomes important in summer (Yanai and Wu 2006; Liu and Wang 2011). However, there is a relative lack of research on how TP heating directly affects the upper WJ (Clark and Serreze 2000; Li *et al.* 2004; Schiemann *et al.* 2008), especially regarding the shift of the WJ from the south to north of the TP during the seasonal spring–summer transition period (Li and Liu 2014). Herein, we analyze the characteristics of the seasonal transition and inter-annual variation of the WJ, and calculate the correlation coefficient of them and the TP heat source based on NCEP reanalysis data. In an attempt to clarify the influence of TP heat source anomaly on the strength and south–north migration of the WJ, three sets of numerical experiments are performed using the latest generation of the Community Atmosphere Model (CAM5) including tests of the TP surface sensible heat flux closure, 50% enhancement, and control.

2. Reanalysis data

This study employs NCEP-DOE reanalysis II (NCEP) and National Oceanic and Atmospheric Administration (NOAA) data (Kanamitsu *et al.* 2002). First, we analyze the daily variation features of the WJ and the relationship between the strength and position of the WJ and TP heat source at the seasonal scale. Second, the correlation coefficients between the westerly wind speed at 200 hPa and the TP surface temperature in spring during the last 35 years are computed. The westerly range in this paper refers to the area within 70°–100°E, 20°–60°N.

2.1 The daily variations of the strength and center position of the WJ

Figure 1(a and b) depicts the average January and July wind fields, respectively, at 200 hPa over the

Figure 1. Westerly wind field (m/s) and WJ axis diurnal variation (°N) at 200 hPa in NCEP for (a) January and (b) July. (c) The diurnal time series of the WJ position and temperature anomalies of high temperature center.

past 35 years. Shading indicates the WJ center position in the two figures. In January, the westerly wind is strong and the maximum wind speed exceeds 40 m/s and is located at approximately 25°N which is to the south of TP. In July, the WJ weakens to 30 m/s and jumps to 40°N which is situated to the north of TP. Overall, the WJ declines in intensity and moves northward with the seasonal transition.

In order to clearly show the features of diurnal variation in the WJ, we calculate the extreme latitude of the WJ axis and zonal temperature anomaly of the high temperature center (80°–100°E, 25°–35°N) at 200 hPa over the TP (figure 1c). The axis of WJ begins its northward movement from 25°N in mid-April, and reaches its northernmost extent at 45°N in mid-July. The westerly changes of the middle and upper troposphere are mainly affected by the temperature gradient (Li *et al.* 2013). The temperature gradient is dependent on the strength of the TP surface heating over the TP in spring. In figure 1(c), the WJ axis and the temperature anomalies rise simultaneously, and this provides additional evidence that the temperature regulates the WJ change. During spring–summer conversion period, the temperature anomalies in the upper tropospheric mainly rely on the TP surface heating.

2.2 Characteristics of TP heat source and WJ in spring during the past 35 years

Figure 2(a and b) depicts the series of the zonal wind south (80°–100°E, 20°–30°N, hereinafter referred as STP) and north (80°–100°E, 40°–50°N, hereinafter referred as NTP) of the TP at 200 hPa as well as the TP surface temperature. The change of the westerly over the NTP is in agreement with the TP surface temperature. The correlation coefficient reaches 0.659 at 99.9% level of significance. In contrast, the correlation coefficient is -0.541 (99.9% significance level) over the STP. There is a negative correlation between the NTP and STP westerly wind. The significant negative correlation is explained by the existence of TP heat source.

The distribution chart of correlation coefficients between the zonal wind and the TP surface temperature is shown in figure 2(c). Adopting TP as the dividing line, there is unquestionably opposite correlations on the north and south sides. Shaded areas represent a significance level of 99.9% (figure 2c). The effects of the TP heat source on the north and southwesterly winds are opposite. The upper atmospheric warming over the TP in spring is mainly composed of the TP surface sensible heating contribution (Li and Yanai 1996). This is an important reason which explains why the TP

Figure 2. The time series of TP surface temperature (K) and 200 hPa zonal winds (m/s) over the south (a) and north (b) of the TP as well as their correlation coefficient distribution (c) in NCEP.

heating to the upper atmosphere causes northward movement of WJ, especially during the spring–summer transition period (Schiemann *et al.* 2008). In spring, the TP surface temperature depends on the TP surface sensible heat flux. Therefore, this paper makes use of the latest generation CAM5 to compare and analyze the test results of suppressed and increased TP surface sensible heat flux, to discuss the effects of TP sensible heating on westerly circulation in spring.

3. Numerical experiments

3.1 Descriptions

The latest generation of CAM5, including the atmosphere and land modules (Neale *et al.* 2010), is a sub-module of the Community Earth System Model (CESM1.0) that was released by the US National Center for Atmospheric Research (NCAR)

in the summer of 2010 (Vertenstein *et al.* 2010). The latest in the series of models developed over several decades, CAM5 has achieved substantial improvement (Liu *et al.* 2012). The most significant improvement is the impact of anthropogenic emissions on the climate in the boundary layer and radiation process, which could not be carefully considered in the previous versions. Since the initial development of Community System Model (CSM), CAM5 has been widely used to model past, present and future climate changes (Meehl *et al.* 2006; Neale *et al.* 2010; Shi and Liu 2011; Gettelman *et al.* 2012; Levis *et al.* 2012; Das *et al.* 2013).

Based on the above analysis, we use the up-to-date version of CAM5 to conduct three sets of experiments to further clarify the main contribution of the TP sensible heat flux to the upper westerly. The first is a control experiment with no altered parameters, referred as CTL. The second

Figure 3. Westerly wind field (m/s) and WJ axis diurnal variation (°N) at 200 hPa in CTL for (a) January and (b) July. (c) The diurnal series of the WJ position and temperature anomalies of high temperature center.

is a TP sensible heat flux lowering experiment (referred to as TPSHL), i.e., suppressing the surface sensible heat flux when the TP (2000 m above sea level) surface sensible heat flux is greater than zero. In this experiment, the TP surface does not transfer sensible heat to the atmosphere in the numerical integration, and the land-atmosphere heat exchange stops (Guo *et al.* 2011; Li *et al.* 2014). The third is a TP sensible heat flux heightening experiment (referred to as TPSHH). In this experiment, the surface sensible heat flux increases by 50% when the TP surface sensible heat flux is greater than 0. In these three numerical experiments, we use the horizontal resolutions of $1.9 \times 2.5^\circ$ with 30 vertical layers. Every experiment integrates over 13 consecutive years, and the average for the last 10 years is employed in the analysis.

Comparing figures 3 and 1, we find that CAM5 successfully simulates the winter and summer westerly circulation and gives reasonable seasonal variations in the strength and center location of the WJ. Simulation results of CAM5 are basically consistent with NCEP reanalysis.

3.2 The results of CAM5

The TP, as an elevated heat source, transports heat to the upper atmosphere through the surface sensible heat flux which results in the formation of a high temperature anomaly center over the TP (Wu *et al.* 2007). During the spring-summer

transition period, the TP sensible heat flux gradually strengthens and upward transport forms a high temperature center at 200 hPa by the end of April (figure 4a), consistent with the reanalysis data (Tamura and Koike 2010). Figure 4 shows only the upward vertical heat flux values for the purpose of displaying the heat transport process. This shows that the high temperature center begins to appear at the beginning of May, which is due to the contribution of the TP surface heating. Previous studies suggest that the Bay of Bengal latent heat release also has a greater contribution to the high temperature center (Tamura *et al.* 2010). Comparing figure 4(a and b), we find that the TP transports almost no heat to the upper atmosphere until the end of May owing to the TP sensible heat flux suppression in TPSHL (figure 4b). Vertically upward heat flux appears again after May in TPSHL modulated by the TP latent heat flux, but the value is smaller than CTL. However, a surge of strong upward vertical heat flux in TPSHH arrives at 200 hPa in early April (figure 4c). The difference between TPSHH and TPSHL demonstrates that the contributions of the TP sensible heat in the troposphere are mainly from April to June, i.e., the spring-summer transition period. These results further illustrate that the experiments in the present paper which modified the TP surface sensible heat flux reasonably represent surface temperature changes.

The spring average temperature difference compared to the same latitude (figure 5a) indicates

Figure 4. Simulated vertical heat flux in CTL, TPSHL and TPSHH and the difference between TPSHH and TPSHL (negative values indicate the upward transport of heat. unit: Pa/s).

Figure 5. Zonal mean temperature anomalies in CTL (a) and NCEP (b) as well as the temperature differences (c and d) between TPSHH, TPSHL and CTL at 200 hPa (unit: K).

that there is an obvious high temperature center at 200 hPa over the TP. This is the same result as in the reanalysis material (figure 5b, Tamura and Koike 2010). The heat center at 200 hPa is located just above the southern of the TP, and this indirectly proves the effect of the TP heat source. In TPSHL, the high temperature anomaly center significantly weakens because of the surface sensible heat flux suppression. There is a key negative difference center between TPSHL and CTL (figure 5c) which corresponds with the lower temperature anomaly over the TP in TPSHL. The maximum cooling exceeds -1.2°C , reaching the 95% significance level (shaded). While in TPSHH, the temperature is generally rising, and the temperature over the TP is higher than the surroundings. However, there is no significant trend at the 95% test level (figure 5d). In summary, we conjecture that TP sensible heat flux contributes significantly to the atmospheric temperature, especially the high temperature center over the TP.

The location of the WJ center is closely related with the north-south temperature gradient. The TP surface sensible heat flux anomalies lead to the tropospheric temperature changes and the WJ center position variations. When the TP sensible heat flux increases 50% in TPSHH, the WJ center position begins moving northward at the end of March. This is nearly two weeks earlier than CTL (figure 6a). In contrast, the WJ center position in TPSHL does not begin to move northward until mid-May, which is a delay of one month compared to CTL. More intense sensible fluxes over the TP lead to a northward movement of the WJ. The largest difference in the WJ center position between TPSHL and CTL is 8° of southerly latitude in mid-May over the TP (figure 6b). The maximum

difference between TPSHH and CTL is 5° of northerly latitude in late May (figure 6b). Therefore, influences of the TP sensible heating on the WJ center situation are apparent.

Similarly, TP sensible heating has a greater impact on the strength of the WJ. The spring 200 hPa wind differences between TPSHL and CTL indicate that there are westerly anomalies over the southern portion of the TP. When the TP surface is cold in spring, the WJ is powerful south of the TP and weak north of the TP (figure 6c). In contrast, there are easterly (westerly) differences south (north) of the TP between TPSHH and CTL in figure 6(d). The shaded areas represent significant level of 95% in figure 6(c and d). In short, the TP heat source influences the position of WJ and also affects the WJ strength.

Generally speaking, the seasonal meridional shift of the WJ can be divided into three periods (Kuang and Zhang 2005): stationary period, northward migration and southward withdrawal. In the present paper, we focus on the northward migration during the spring-summer transition period. Figure 7 depicts the simulated monthly location of the WJ from April to August. For the CTL test (figure 7a), the WJ is steadily located near 30°N in April, and then it moves northward by approximately 5° of latitude from April to May. The amplitude of the shift over the TP is larger than the surrounding of the TP. The first integral northward jump of the WJ is consistent with the onset of South China Sea summer monsoon (Li *et al.* 2004; Kuang and Zhang 2005). After May, the WJ gradually moves northward and reaches the northernmost position in August. In the TPSHL, where the sensible heat flux is suppressed over the TP, the WJ is still steadily situated near 30°N

Figure 6. The WJ axis locations ($^{\circ}$ N) and the differences in wind field (m/s): (a) WJ axis locations in CTL, TPSHL and TPSHH; (b) differences of WJ position between TPSHL, TPSHH and CTL; (c) 200 hPa wind field difference between TPSHL and CTL in spring and (d) 200 hPa wind field difference between TPSHH and CTL.

Figure 7. The seasonal evolution of the WJ axis from April to August (a) CTL, (b) TPSHL and (c) TPSHH.

in May as well as April, i.e., the first northward jump occurs from May to June (figure 7b). In contrast, the WJ in TPSHH moves northward by

approximately 10° of latitude from April to May. This is 5° larger than in CTL and was caused by the 50% enhancement of sensible heat flux (figure 7c). Similarly, the amplitude of the shift over TP is smaller (larger) than the surrounding of the TP in TPSHL (TPSHH). In summary, there are obvious association between the northward migration of the WJ and the sensible heat flux of the TP from April to June.

4. Conclusion and discussion

This study investigates the last 35 years of NCEP data and our analysis suggests that the intensity and location of the WJ are closely linked with the TP heat source. The position of WJ keeps pace with the temperature gradient changes at 200 hPa. The WJ is located south of the TP in winter, and jumps to the north of the TP in summer. During the last 35 years, the correlation coefficient between the zonal wind over the south (north) of the TP and the TP surface temperature in spring is -0.541 (0.659) with a significance level of 99.9%.

CAM5 reasonably simulates the seasonal variation in the strength and location of the WJ and is consistent with the NCEP conclusion. The results from the suppressed and enhanced TP sensible heat flux sensitivity experiments suggest that the TP heat source determines the WJ strength and position changes during the spring–summer transition period. When suppressing the TP sensible heat in TPSHL, no vertical conveyance of upward heat until the end of May causes upper troposphere cooling -1.2°C with a significance level of

95%. This leads to an enhancement and southward movement of the WJ. The first northward jump of the WJ in TPSHL occurs from May to June which is one month delayed relative to CTL. The maximum gap of center location of the WJ between TPSHL and CTL is 8° of southerly latitude which appears in mid-May. The difference between TPSHL and CTL in the wind field at 200 hPa indicates that there is westerly (easterly) deviation over the south (north) of TP. However, increasing the TP sensible heat flux by 50% results in opposite changes. The largest difference in the position of the WJ axis between TPSHL and CTL is 5° of latitude. In the experiment with the enhancement of the sensible heat flux over the TP, the WJ moves northward by approximately 10° of latitude from April to May and results in the formation of an easterly (westerly) wind anomaly over the south (north) of the TP with a significance level of 95%.

Both results of reanalysis data and CAM5 indicate significant correlations between the 200 hPa zonal wind and TP heat source. Our conclusion is focused on the seasonal and inter-annual variations of the WJ. There is a critical need to evaluate our results over longer timescales. Observational and simulational studies show that the TP significantly warms and exhibits an early signal with global climate warming (Kattel and Yao 2013). The warming characteristics of TP are expected to continue in models of future climate change (Liu *et al.* 2009). This result predicts that with the future global climate warming, the WJ will weaken and exhibit an early move northward modulated by TP heat source in spring. Therefore, considerable attention must be paid to the location and strength of the WJ with respect to the sensible heat flux over the TP. In future studies, we will quantitatively assess the response of the WJ to the TP heat flux and focus on the evolution features of the WJ response to different future emission scenarios.

Acknowledgements

The authors thank the editor and anonymous reviewer for their constructive comments that helped to improve this manuscript. This study was sponsored by the National Natural Science Foundation of China (Grant nos. 41472162 and 41290255) and the Strategic Priority Research Program of the Chinese Academy of Sciences (Grant no. XDB03020601).

References

- Chen L X 1957 The variation of wind field along 140°E at the end of Meiyu; *Acta Meteor. Sin.* **28**(4) 294–302.
- Chen B D, Chao W C and Liu X D 2003 Enhanced climatic warming in the Tibetan Plateau due to double CO₂: A model study; *Clim. Dyn.* **20** 401–413.
- Clark M P and Serreze M C 2000 Effects of variations in East Asian snow cover on modulating atmospheric circulation over the North Pacific Ocean; *J. Climate* **13** 3700–3710.
- Das S K, Deb S K, Kishtawal C M and Pal P K 2013 Seasonal prediction of Indian summer monsoon: Sensitivity to persistent SST; *J. Earth Syst. Sci.* **122**(5) 1183–1193.
- Ding Y H 1992 Effect of the Qinghai–Xizang (Tibetan) Plateau on the circulation features over the plateau and its surrounding areas; *Adv. Atmos. Sci.* **9** 112–130.
- Duan A M, Liu Y M and Wu G X 2005 Heating status of the Tibetan Plateau from April to June and rainfall and atmospheric circulation anomaly over east Asia in mid-summer; *Sci. China Earth Sci.* **48**(2) 250–257.
- Feng S, Tang M C and Wang D M 1998 New evidence for the Qinghai–Xizang (Tibet) Plateau as a pilot region of climatic fluctuation in China; *Chinese Sci. Bull.* **43**(20) 1745–1749 (in Chinese).
- Flohn H 1957 Large-scale aspects of the ‘summer monsoon’ in south and east Asia; *J. Meteorol. Soc. Japan* **35** 180–186.
- Gettelman A, Kay J E and Shell K M 2012 The evolution of climate sensitivity and climate feedbacks in the community atmosphere model; *J. Climate* **25** 1453–1469.
- Guo X F, Yang K and Chen Y Y 2011 Weakening sensible heat source over the Tibetan Plateau revisited: Effects of the land–atmosphere thermal coupling; *Theor. Appl. Climatol.* **104** 1–12.
- Kanamitsu M, Wesley E, Woolen J, Yang S K, Hnilo J J, Fiorino M and Potter G L 2002 NCEP–DOE AMIP–II Reanalysis (R-2); *B. Am. Meteorol. Soc.* **83** 1631–1643.
- Kattel D B and Yao T D 2013 Recent temperature trends at mountain stations on the southern slope of the central Himalayas; *J. Earth Syst. Sci.* **122** 215–227.
- Krishnamurti T N 1961 The subtropical jet stream of winter; *J. Meteorol.* **18** 172–191.
- Kuang X Y and Zhang Y C 2005 Seasonal variation of the east Asian subtropical westerly jet and its association with the heating field over east Asia; *Adv. Atmos. Sci.* **22** 831–840.
- Levis S, Bonan G, Kluzek E, Thornton P E, Jones A, Sacks W J and Kucharik C 2012 Interactive crop management in the Community Earth System Model (CESM1): Seasonal influences on land–atmosphere fluxes; *J. Climate* **25** 4839–4859.
- Li X Z and Liu X D 2014 Numerical simulation of the impact of spring sensible heat anomalies over Tibetan Plateau on rainy season precipitation in north China; *Chinese J. Earth Env.* **5**(3) 207–215.
- Li C and Yanai M 1996 The onset and interannual variability of the Asian summer monsoon in relation to land–sea thermal contrast; *J. Climate* **9** 358–375.
- Li C Y, Wang Z T, Lin S Z and Cho H R 2004 The relationship between east Asian summer monsoon activity and northward jump of the upper westerly jet location; *Chinese J. Atmos. Sci.* **28**(5) 641–658.
- Li X Z, Liu X D, Qiu L J, An Z S and Yin Z Y 2013 Transient simulation of orbital-scale precipitation variation in monsoonal east Asia and arid central Asia during the past 150 ka; *J. Geophys. Res.* **118** 7481–7488.
- Li Y, Wang N A, Zhou X H, Zhang C Q and Wang Y 2014 Synchronous or asynchronous Holocene Indian and east Asian summer monsoon evolution: A synthesis on Holocene Asian summer monsoon simulation, records and modern monsoon indices; *Global Planet. Change* **116** 30–40.

- Liu X D and Chen B D 2000 Climatic warming in the Tibetan Plateau during recent decades; *Int. J. Climatol.* **20** 1729–1742.
- Liu X D and Wang Y 2011 Contrasting impacts of spring thermal conditions over Tibetan Plateau on late-spring to early-summer precipitation in southeast China; *Atmos. Sci. Lett.* **12**(3) 309–315.
- Liu X, Easter R C and Ghan S J 2012 Toward a minimal representation of aerosols in climate models: Description and evaluation in the Community Atmosphere Model CAM5; *Geosci. Model Dev.* **5** 709–739.
- Liu X D, Cheng Z G, Yand L B and Yin Z Y 2009 Elevation dependency of recent and future minimum surface temperature trends in the Tibetan Plateau and its surroundings; *Global Planet. Change* **68** 164–174.
- Luo H B and Yanai M 1984 The large-scale circulation and heat sources over the Tibetan Plateau and surrounding areas during the early summer of 1979. Part I: Precipitation and kinematic analyses; *Mon. Wea. Rev.* **112** 966–989.
- Matsumoto J 1992 The seasonal changes in Asian and Australian monsoon region; *J. Meteor. Soc. Japan* **70** 257–273.
- Meehl G A, Washington W M, Santer B D, Collins W D, Arblaster J M, Hu A, Lawrence D M, Teng H, Buja L E and Strand W G 2006 Climate change projections for the twenty-first century and climate change commitment in the CCSM3; *J. Climate* **19** 2597–2616.
- Mölg T, Maussion F and Scherer D 2013 Mid-latitude westerlies as a driver of glacier variability in monsoonal high Asia; *Nat. Clim. Change* **4** 68–73.
- Nagashima K, Tada R and Toyoda S 2013 Westerly jet–east Asian summer monsoon connection during the Holocene; *Geochim. Geophys. Geosyst.* **14**(12) 5041–5053.
- Neale R B, Chen C C and Gettelman A *et al.* 2010 Description of the NCAR Community Atmosphere Model (CAM 5.0). NCAR Technical Note NCAR/TN-XXX+STR, National Center for Atmospheric Research, Boulder, Colorado, draft, available from <http://www.cesm.ucar.edu/models/cesm1.0/cam/>.
- Pepin N R, Bradley S, Diaz H F, Baraer M and Caceres B *et al.* 2015 Elevation-dependent warming in mountain regions of the world; *Nat. Clim. Change* **5** 424–430.
- Porter S C and An Z S 1995 Correlation between climate events in the North Atlantic and China during the last glaciation; *Nature* **375** 305–308.
- Sampe T and Xie S P 2010 Large-scale dynamics of the Meiyu–Baiu rainband: Environmental forcing by the westerly jet; *J. Climate* **23** 113–134.
- Schiemann R, Lüthi D and Schär C 2008 Seasonality and interannual variability of the westerly jet in the Tibetan Plateau region; *J. Climate* **22**(11) 2940–2957.
- Shi Z G and Liu X D 2011 Distinguishing the provenance of fine-grained eolian dust over the Chinese Loess Plateau from a modeling perspective; *Tellus B* **63** 959–970.
- Sun Y B, Clemens S C, Morrill C, Lin X P, Wang X L and An Z S 2012 Influence of Atlantic meridional overturning circulation on the east Asian winter monsoon; *Nat. Geosci.* **5** 46–49.
- Tamura T and Koike T 2010 Role of convective heating in the seasonal evolution of the Asian summer monsoon; *J. Geophys. Res.* **115** D14103, doi: 10.1029/2009JD013418.
- Tamura T, Taniguchi K and Koike T 2010 Mechanism of upper tropospheric warming around the Tibetan Plateau at the onset phase of the Asian summer monsoon; *J. Geophys. Res.* **115** D02106, doi: 10.1029/2008JD011678.
- Versteinsten M, Craig T, Middleton A, Feddema D and Fischer C 2010 CESM1.0.3 User's Guide, available at: <http://www.cesm.ucar.edu/models/cesm1.0/cesm/cesmdoc/book1.html>.
- Wu G X and Zhang Y S 1998 Tibetan plateau forcing and the timing of the monsoon onset over south Asia and the south China sea; *Mon. Wea. Rev.* **126** 913–927.
- Wu G X, Liu Y M, Wang T M, Wan R J, Liu X, Li W P, Wang Z Z, Zhang Q, Duan A M and Liang X Y 2007 The influence of mechanical and thermal forcing by the Tibetan Plateau on Asian climate; *J. Hydrometeorol.* **8** 770–789.
- Yanai M, Li C F and Song Z S 1992 Seasonal heating of the Tibetan Plateau and its effects of the evolution of the Asian summer monsoon; *J. Meteorol. Soc. Japan* **70** 319–352.
- Yanai M and Wu G X 2006 Effects of the Tibetan Plateau; chapter 13, In: *The Asian Monsoon* (ed.) Wang B (Berlin, Germany: Springer), pp. 513–549.
- Yang S, Lau K M and Kim K M 2002 Variations of the east Asian jet stream and Asian–Pacific–American winter climate anomalies; *J. Climate* **15** 306–325.
- Yang K, Wu H, Qin J, Lin C G, Tang W J and Chen Y Y 2014 Recent climate changes over the Tibetan Plateau and their impacts on energy and water cycle: A review; *Global Planet. Change* **112** 79–91.
- Yao T D, Thompson L, Yang W, Yu W, Gao Y, Guo X, Yang X, Duan K, Zhao H, Xu B, Pu J, Lu A, Xiang Y, Kattel D B and Joswiak D 2012 Different glacier status with atmospheric circulations in Tibetan Plateau and surroundings; *Nat. Clim. Change* **2**(9) 663–667.
- Ye D Z and Gao Y X 1979 The Meteorology of the Qinghai-Xizang (Tibet) Plateau (in Chinese); Science Press, 278p.
- Ye D Z, Luo S W and Zhu B Z 1957 The wind structure and heat balance in the lower troposphere over Tibetan Plateau and its surroundings; *Acta Meteor. Sin.* **2**(2) 108–121.
- Yu R C, Wang B and Zhou T J 2004 Tropospheric cooling and summer monsoon weakening trend over east Asia; *Geophys. Res. Lett.* **31** L22212, doi: 10.1029/2004GL021270.
- Zhang X Y, Arimoto R and An Z S 1997 Dust emission from Chinese desert sources linked to variations in atmospheric circulation; *J. Geophys. Res.* **102** 28,041–28,047.
- Zhang Y C, Kuang X Y, Guo W D and Zhou T J 2006 Seasonal evolution of the upper-tropospheric westerly jet core over east Asia; *Geophys. Res. Lett.* **33** L11708, doi: 10.1029/2006GL026377.
- Zhao P, Zhou Z and Liu J 2007 Variability of Tibetan spring snow and its associations with the hemispheric extratropical circulation and east Asian summer monsoon rainfall: An observational investigation; *J. Climate* **20** 3942–3955.
- Zhou X J, Zhao P, Chen J M, Chen L X and Li W L 2009 Impacts of thermodynamic processes over the Tibetan Plateau on the Northern Hemispheric climate; *Sci. China Earth Sci.* **52**(11) 1679–1693.