On the Climatology and Trend of the Atmospheric Heat Source over the Tibetan Plateau: An Experiments-Supported Revisit

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ABSTRACT

Atmospheric heating over the Tibetan Plateau (TP) enhances the Asian summer monsoon. This study presents a state-of-the-art estimate of the heating components and their total over the TP, with the aid of high-accuracy experimental data, an updated land surface model, and carefully selected satellite data.

The new estimate differs from previous estimates in three aspects: 1) different seasonality—the new estimation shows the maximum total heat source occurs in July (the mature period of the monsoon), rather than in the previously reported month of May or June (around the onset of the monsoon), because previous studies greatly overestimated radiative cooling during the monsoon season [June–August (JJA)]; 2) different regional pattern—the eastern TP exhibits stronger heating than the western TP in summer, whereas previous studies gave either an opposite spatial pattern because of overestimated sensible heat flux over the western TP or an overall weaker heat source because of overestimated radiative cooling; and 3) different trend—sensible heat, radiative convergence, and the total heat source have decreased since the 1980s, but their weakening trends were overestimated in a recent study. These biases in previous studies are due to fairly empirical methods and data that were not evaluated against experimental data.

1. Introduction

The Tibetan Plateau (TP) exerts a huge heat source on the atmosphere in the Northern Hemisphere during boreal spring and summer (Flohn 1957; Ye and Gao 1979). Because the average elevation of the TP is more than 4000 m MSL and thus the total mass of the air column over the TP is much less than that over its surroundings, the atmospheric heating is more efficient in this region. Furthermore, the sensible heat from the TP ground can directly warm the midtroposphere. It has been indicated that the heating over the TP not only enhances the Asian monsoon circulation (He et al. 1987; Yanai et al. 1992; Wu and Zhang 1998; Qian et al. 2004; Sato and Kimura 2007) but also triggers vigorous deep convections over the TP (Yang et al. 2004) that greatly enhance the troposphere–stratosphere exchanges of water vapor and air pollutants (Fu et al. 2006). Zhou et al. (2009) even suggested that the thermodynamic processes over the TP may impact the climate at a hemispheric scale and even a global scale through modulating atmosphere–ocean interactions.
Meanwhile, the TP exhibits a striking warming since the 1980s, which may have influenced rainfall in east China (Wang et al. 2008) and interacted with the weakening of the Asian monsoon in recent decades (Duan and Wu 2008, hereafter DW08). Recent field experiments on the TP have advanced our understanding of the heating processes from the atmospheric boundary layer up to the free atmosphere (Yang et al. 2004; Taniguichi and Koike 2007; Tamura et al. 2010), but how to quantify the heat source and its trend is still a major topic of TP meteorology. A high-accuracy estimate of the heat source over the TP is a basis for testing the capability of a climate model or a reanalysis system to reproduce the atmospheric energy cycle in the Asian monsoon system. Such an estimate also contributes to setting reasonable conditions (e.g., surface sensible heat flux, diabatic heating) for process studies by numerical modeling (e.g., Kuwagata et al. 2001; Liu et al. 2007; Wu et al. 2007).

The total heat source (TH) comprises three components, that is, surface sensible heat (SH), latent heat release of condensation (LH), and radiative convergence (RC). In the literature, there are two approaches to estimate the heat source.

The first approach is to vertically integrate the local change rate and advection of energy derived from a reanalysis dataset. The accuracy of this approach highly depends on the amount of observations assimilated into an analysis model. The TP has an area of about $2.5 \times 10^6$ km$^2$, but there are only a few operational radiosonde stations whose data have been assimilated into current reanalysis products. Therefore, a high-accuracy reanalysis dataset is still absent, except for several episodes, such as the First Global Atmospheric Research Program (GARP) Global Experiment (FGGE) during the summer of 1979 and the Global Energy and Water Cycle Experiment (GEWEX) Asian Monsoon Experiment–Tibet (GAME–Tibet) during the summer of 1998. Heat source estimations using these episode data provided insights into the relationship between the atmospheric heating over the TP and the summer monsoon’s seasonal march (Luo and Yanai 1984; Yanai et al. 1992; Ueda et al. 2003); however, they are not applicable of long-term investigations, such as interannual variability and climatic change of the heat source.

The second approach to estimate the heat source is based on surface data and/or satellite products. This type of estimation was first presented in Ye and Gao (1979) and followed by many studies (e.g., Chen et al. 1985; Zhao and Chen 2000b; Li and Chen 2003). Among various components, SH and LH were estimated from the China Meteorological Administration (CMA) station data. However, shortcomings are obvious in these studies. First, SH was estimated by the bulk method with a constant heat transfer coefficient or a simple parameterization. Yang et al. (2011) pointed out these empirical methods may produce opposite results in SH trend, as they do not follow fundamental micrometeorological theory. Second, the term of radiative convergence was estimated from satellite products without recognizing their biases. Yang et al. (2006a) called attention to the accuracy of current satellite estimates of radiation budget. Particularly, they may have considerable biases for the TP region because of its extreme climatic conditions. Third, the CMA stations are irregularly distributed, but their representativeness for use of estimating the heat source has not been clarified. Therefore, large uncertainties may exist in previous estimates because of limitations in both data availability and methods. These issues will be addressed in this study.

As shown later, previous estimates of the heat source gave rather mixed results. Meanwhile, several hydrometeorological experiments have been implemented on the TP since the end of the 1990s. The obtained data have significantly advanced our understanding of land–atmosphere interaction processes, and they have been applied to land model development (Yang et al. 2009a; van der Velde et al. 2009; Chen et al. 2010). Although these experimental data have been available for nearly 10 years, they were not used for estimating the heat source in previous studies.

The objective of this study is to provide a state-of-the-art estimate of the TP heat source terms and their trends during recent decades. In contrast to previous estimates, the new estimate relies on recent advances in field data analyses, land surface modeling, and satellite products’ evaluation. The heat source was estimated for 3 TP sub-regions with distinct climate: dry western TP or W–TP (west of $85^\circ$E), transitional central TP or C–TP ($85^\circ$–$95^\circ$E), and wet eastern TP or E–TP (east of $95^\circ$E).

The paper is organized as follows. Three types of data are first introduced (section 2) followed by a new method and a conventional method to quantify the components of the heat source (section 3). Then, the climatology of the estimated heat source terms and their trends are presented and compared with a conventional empirical estimate (sections 4 and 5, respectively). The representativeness of the heat source terms estimated from sparsely distributed stations are discussed in section 6, and concluding remarks are given in section 7.

2. Data

a. Satellite data

Two satellite products of surface radiation budget and top of the atmosphere (TOA) radiation budget were used in this study. They are the GEWEX Surface Radiation Budget (GEWEX SRB; Stackhouse et al. 2004) and the
International Satellite Cloud Climatology Project Flux Data (ISCCP–FD; Zhang et al. 2004). The former was estimated with the cloud cover and radiances from the ISCCP stage DX (ISCCP–DX) nominal 30-km pixels within each 1° × 1° cell. The latter was estimated with the ISCCP, stage D1 (ISCCP-D1) cloud cover, cloud-top temperature, optical thickness, and cloud phase based on 15 cloud types in a 280-km (2.5°) equal-area grid with climatologies for cloud particle size and vertical structure. The GEWEX SRB product has several versions, and this study used version 3.0 for shortwave radiation (SW) and version 2.0 for longwave radiation (LW).

Another satellite product involved in this study is the Tropical Rainfall Measuring Mission (TRMM) 3B42 (V6) precipitation, with a 3-h temporal resolution and a 0.25° spatial resolution in a global belt spanning 50°S–50°N in latitude. This product is a fusion of microwave precipitation estimates and infrared precipitation estimates, which was then scaled to match monthly rain gauge analyses (Huffman et al. 2007).

The radiation products are available since 1984, and they are used to estimate RC and evaluate the spatial representativeness of station data. The TRMM precipitation data are available since 1998, which are used to upscale LH from stations to a regional scale.

b. CMA routine data

CMA provides long-term station data for this study. Figure 1 shows the distribution of CMA stations on the TP (marked by solid circles in the figure). There are a total of 85 stations. Each station started operations in different years, and some measurements are not available at the early stage. There are a total of 21 stations that have “good data records” (defined as data records that are available for more than 300 days of a year) since 1961; however, the number of the stations was increased to 78 during 1984–2006, of which there are 3, 20, and 55 on the W-TP, C-TP, and E-TP, respectively. The simultaneous availability of both satellite data and the CMA dataset makes this study focus on the heat source estimation during 1984–2006.

At each CMA station, measured meteorological parameters are 6-hourly [0200, 0800, 1400, 2000 Beijing standard time (BST); BST = UTC + 8 h] wind speed, air temperature, and specific humidity; maximum, minimum, and daily mean air pressure; daily precipitation; and daily sunshine duration. Instead of being directly used to estimate sensible heat flux in many previous studies, these observed data are used herein to drive a land surface model (LSM) that produces the surface net radiation and sensible heat flux.

Figure 2 shows the observed trends in annual-mean wind speed and air temperature in the subregions. It is shown that the wind speed has significantly decreased, which is a common phenomenon in China (Xu et al. 2006), while the air temperature increased during recent decades. More details of the temperature trend were analyzed with satellite data (Oku et al. 2006; Qin et al. 2009) and reanalysis data (You et al. 2010). The response of SH to the TP climate changes was investigated in this study.

c. Tibetan experimental data

The GAME–Tibet in 1998 and the follow-on post-GAME and Coordinated Enhanced Observation Period
(CEOP) experiments during 1999–2004 collected high-accuracy data for TP hydrometeorological studies (Koike et al. 1999; Ueno et al. 2001; Ma et al. 2002b; Tanaka et al. 2003). Among these experimental stations, four automatic weather stations (AWSs) were colocated with CMA stations, and thus they were selected to evaluate land surface modeling results and satellite products in this study. The AWSs are marked by a circle within a circle in Fig. 1. Two stations were deployed on the W-TP [Shiquanhe (SQH) and Gerze (or Gaize)] and the other two stations on the C–TP (Naqu and Anduo). The measuring periods at the experimental sites are different: May–September 1998 for SQH, May–June 1998 and October 2002–December 2004 for Gerze, June–September 1998 and October 2002–August 2004 for Naqu, and June 1998–May 2003 and May–December 2004 for Anduo. The experimental data are used to evaluate heat source estimates.

3. Method description

A new estimation and a typical conventional method were presented and compared to each other. The conventional method is very similar to the one presented in DW08. Herein, the phrase “conventional method” means that the used data and method are either empirical or not evaluated against experimental data. Table 1 summarizes the differences between the two methods and the details are presented below.

a. Estimating sensible heat flux

1) CONVENTIONAL METHOD

Conventionally, SH was estimated from CMA station data by the following bulk method:

$$SH = c_p \rho C_H u (T_g - T_a),$$  \hspace{1cm} (1)

where $\rho$ (kg m$^{-3}$) is the air density, $c_p$ (1004 J kg$^{-1}$ K$^{-1}$) is the specific heat of air at constant pressure, $u$ (m s$^{-1}$) is the observed wind speed at level $z_m$ (m), $T_g$ (K) is the observed ground temperature, $T_a$ (K) is the observed air temperature at level $z_h$ (m), and $C_H$ is the bulk heat transfer coefficient.

There are a number of studies on how to determine $C_H$ value for the TP. A brief review of these studies is given in Yang et al. (2011). Similar to DW08, $C_H$ was set to be 0.004 for the C–TP and the E–TP stations, while it was set to 0.00475 for the W-TP stations. The CMA 6-hourly data are used as input for Eq. (1). DW08 assumed $\rho = 0.8$ kg m$^{-3}$ in Eq. (1), but it was calculated from observed air temperature and air pressure in this study. This difference does not fundamentally alter the climatology of the sensible heat flux and its trend.

<table>
<thead>
<tr>
<th>Heat source component</th>
<th>Conventional</th>
<th>New</th>
</tr>
</thead>
<tbody>
<tr>
<td>SH</td>
<td>Empirical bulk equation</td>
<td>Improved LSM simulation</td>
</tr>
<tr>
<td>LH</td>
<td>Estimated from CMA precipitation data</td>
<td>Estimated from corrected CMA precipitation data</td>
</tr>
<tr>
<td>RC</td>
<td>Estimated from ISCCP-FD data</td>
<td>Estimated from selected satellite data and LSM simulation</td>
</tr>
</tbody>
</table>
2) NEW METHOD

Xu and Haginoya (2001) showed that it is possible to produce reasonable SH values on the TP region by land surface modeling. In the new estimate, SH was simulated by the second-generation Simple Biosphere model (SiB2; Sellers et al. 1996), which has been improved according to experimental data analyses and TP surface conditions (Yang et al. 2009a). The improvements include the implementation of an aerodynamic canopy model suitable for short and sparse vegetation (Watanabe and Kondo 1990), a new surface flux parameterization scheme developed from the experimental data (Yang et al. 2008), a high-accuracy soil water flow scheme (Ross 2003) with additional considerations of soil freezing and thawing, and a new parameterization of soil surface evaporation resistance. The improved model performs better than the original one in simulating surface temperature, sensible heat flux, and net radiation (Yang et al. 2009a).

The simulation requires input of high-resolution data of wind speed, air temperature, specific humidity, downward radiation, and precipitation. They were downscaled from CMA 6-hourly or daily data. A statistical method to disaggregate wind speed and air temperature from CMA 6-hourly data to half-hourly data was developed in Yang et al. (2009b, hereafter Yang09), and it was adopted herein. Specific humidity was linearly interpolated from CMA 6-hourly data. Daily precipitation was uniformly distributed for the hours from late afternoon to early morning, as suggested from high-resolution experimental data (Yang et al. 2007). Downward SW was not observed at most of the CMA stations and LW was totally not observed. In this study, SW was estimated from CMA sunshine data by the model in Yang et al. (2006b, hereafter Yang06) and LW estimated by the scheme presented in Crawford and Duchon (1999, hereafter CD99). This CMA data-based estimate of downward SW and LW is more accurate than the two satellite estimates (Fig. 3), and thus it was used to drive the land surface modeling.

Relevant soil parameters and vegetation parameters (classification and coverage) in the model were derived from 1° × 1° International Satellite Land Surface Climatology Project Initiative II (ISLSCP II) soil data (Global Soil Data Task Group 2000) and vegetation data (Loveland et al. 2001). Leaf area index data were sourced from Moderate Resolution Imaging Spectroradiometer...
(MODIS) 0.25° × 0.25° gridded 8-day leaf area index products (Knyazikhin et al. 1999).

To achieve a sufficient modeling spinup for dry stations, the simulated period started from 1952. Climatologic data were used to drive the model until the period for which observations are available.

b. Estimating latent heat release of condensation

Latent heat release due to water vapor condensation can be estimated as follows:

\[ LH = L \rho_w P \]  

(2)

where \( \rho_w \) is the density of water (kg m\(^{-3}\)), \( P \) (mm s\(^{-1}\)) is the surface precipitation, and \( L \) is the specific heat of evaporation or sublimation (J kg\(^{-1}\)).

At CMA stations, precipitation was measured with the Chinese standard precipitation gauge since the late 1950s. The gauge is a cylinder of galvanized iron, 65 cm high and 20 cm in diameter. It was placed 0.7 m above the ground without a windshield. According to Ye et al. (2004), a correction to CMA gauge-measured precipitation was made to account for wind-induced error, wetting loss, and trace precipitation. This correction was made once a day and the corrected amount is about 15% of the gauged precipitation for the wetter C–TP and E–TP and about 30% for the drier W–TP. As this correction depends on wind speed and air temperature, which has changed in recent decades, the trend in precipitation is slightly altered by the correction (Ding et al. 2007). Nevertheless, this change is not very significant and not concerned further.

In Eq. (2), the conventional method uses the CMA gauge-measured precipitation, whereas the new method uses the corrected precipitation.

c. Estimating radiative convergence

Radiative convergence was calculated as follows:

\[ RC = R_{\text{toa}} - R_{\text{sfc}} \]  

(3)

where \( R_{\text{toa}} \) and \( R_{\text{sfc}} \) are net radiation fluxes at TOA and at the ground surface, respectively.

Both GEWEX SRB and ISCCP–FD provide the data for \( R_{\text{toa}} \) and \( R_{\text{sfc}} \). The land surface modeling also yields \( R_{\text{sfc}} \). Zhao and Chen (2000a) presented an empirical method to estimate \( R_{\text{toa}} \) and \( R_{\text{sfc}} \). Ma et al. (2002a) developed another remote sensing approach to estimate \( R_{\text{sfc}} \), but it is not applicable to cloudy conditions. Attention needs to be paid to the use of these datasets. For instance, Fig. 3 shows that ISCCP–FD SW and LW data are quite biased, though it was used in DW08. In Zhao and Chen (2000a), the estimated wintertime \( R_{\text{sfc}} \) is comparable to observations over the W–TP, but the summertime one is about 20 W m\(^{-2}\) higher than the observations (not shown). Therefore, a dataset should be evaluated before being used for the RC estimation.

Figure 4a shows the comparison between annual mean \( R_{\text{toa}} \) of GEWEX SRB and ISCCP–FD after averaged over three subregions. It is seen that the two satellite estimates give comparable \( R_{\text{toa}} \) trends. By contrast, the two satellite estimates and the LSM estimate of \( R_{\text{sfc}} \) are very different; their discrepancies in annual-mean values can be up to 30 W m\(^{-2}\) (Fig. 4b). The significant negative trend in GEWEX SRB \( R_{\text{sfc}} \) observed in Fig. 4b is mainly due to a strong negative trend in outgoing (reflected) shortwave radiation (0.55 W m\(^{-2}\) yr\(^{-1}\)), while the large interannual variability in ISCCP–FD \( R_{\text{sfc}} \) is due to the large interannual variability of all radiation components, except for incoming shortwave radiation (not shown). The comparable \( R_{\text{toa}} \) trends and the mixed \( R_{\text{sfc}} \) trends are not surprising, as it is more difficult to retrieve surface radiation budget than to retrieve TOA radiation budget.

To select the least error-prone \( R_{\text{sfc}} \), Fig. 5 compares the observed and the estimated monthly-mean \( R_{\text{sfc}} \) at the four Tibetan experimental sites. Clearly, the two satellite...
products overestimated $R_{n_{dsc}}$, but the LSM-simulated $R_{n_{dsc}}$ is closer to the observations.

To reduce the uncertainties, RC in the new estimation was calculated with $R_{n_{dsc}}$ from the land surface modeling and $R_{n_{toa}}$ averaged over the two satellite products. By contrast, the conventional estimate was estimated from ISCCP-FD radiation data, as adopted in DW08.

d. Estimating total heat source

The total heat source is the arithmetic sum of the above three components as follows:

$$TH = SH + LH + RC.$$  \hfill (4)

4. Heat source climatology over the Tibetan Plateau

Figure 6 compares the climatology of monthly-mean heat source terms between the conventional estimate and the new estimate in each subregion. Figure 7 compares the new estimate with several previous estimates for the entire plateau. Attention needs to paid to the periods and averaging methods used for Fig. 7. The study periods are different: before 1979 in Ye and Gao (1979), a single year of 1979 in Yanai et al. (1992), 1961–95 in Zhao and Chen (2001), and 1984–2006 in this study. The methods to average the heat source terms from the subregions to the entire plateau are also different among these studies. These differences may contaminate the comparison but do not fundamentally alter our conclusions.

a. Sensible heat flux

Figure 6a shows that some basic features of the two SH estimates in this study are similar to each other. The maximum SH in each subregion roughly occurs in May and the minimum SH occurs in December. SH decreases from the W–TP to the E–TP, and this regional contrast is associated with regional differences in rainfall and the surface feature (Fig. 1). Nevertheless, the new SH’s magnitude in Fig. 6a is generally less than the conventional one, and their differences are particularly large over the W–TP. The higher accuracy of the new estimate can be justified by the following analysis of the surface energy budget over the W–TP at both annual and monthly scales.

According to experimental data at Gerze station on the W–TP, the annual-mean surface net radiation averaged over 2003/04 is 76 W m$^{-2}$. At annual scale, ground heat flux ($G_0$) is near 0; thus, the net radiation ($R_{n_{dsc}}$) is primarily partitioned into sensible heat (SH) and latent heat (LE). The W–TP is dry, with annual precipitation of about 160 mm; thus, precipitated water is almost all lost through evaporation because of very strong solar radiation. This leads to annual-mean LE of 13 W m$^{-2}$; thus, annual-mean SH should be about 63 W m$^{-2}$. Based on experimental data in 1998, Li et al. (2000) obtained SH of 67 W m$^{-2}$ for this station. These SH values are close to the new value (62 W m$^{-2}$) but less than the conventional one (76 W m$^{-2}$).

The maximum SH occurred in May and June. According to observations at Gerze station, the monthly-mean $R_{n_{dsc}}$ averaged over 2003/04 is 110 W m$^{-2}$ in May and 129 W m$^{-2}$ in June. At the Shiquanhe experimental site, the net radiation is even less. As the monthly mean of $G_0$ and LE are about 10 W m$^{-2}$, it is generally not expected that monthly-mean SH on the W–TP is larger than 100 W m$^{-2}$. Zhao and Chen (2001), according to the surface energy budget and observational data, gave a SH value of 100 W m$^{-2}$ for May and June. This value is close to our new estimate (98 W m$^{-2}$), while it is much less than the conventional estimate for the two months (~140 W m$^{-2}$). In reality, most of previous studies, using either reanalysis data or CMA data, greatly overestimated SH in May and June (ranging from 120 to 220 W m$^{-2}$ on the W–TP), as summarized in Table 2. On the relatively wet C–TP and E–TP, maximum SH should be less than the one on the W–TP; therefore, the values for the C–TP and the E–TP in Table 2 are also too high to trust.
Figure 7a compares SH between the new estimate and the two estimates in the literature for the entire plateau. It is observed that the estimate of Zhao and Chen (2001), which is approximately constrained by the surface energy budget, is close to the new estimate, whereas an earlier effort by Ye and Gao (1979) greatly overestimated SH because of a large value (0.008) of $C_{T H}$ being used.

In brief, SH estimates in most previous studies were too high for May and June, a period of major concern in monsoon studies, because of their omission of the physical constraint by the surface energy budget.

**b. Latent heat release of condensation**

Figure 6b shows the monthly-mean LH climatology. The two methods produced a clear seasonal variation of LH. The maximum LH in each subregion occurs about July–August, and the minimum LH occurs in December–January. In contrast to SH, LH drastically increases from...
the W–TP to the E–TP, as the plateau monsoon migrates from the southeast to the northwest. The correction to CMA precipitation gives slightly higher LH values over the W–TP and during the non monsoon seasons as well, whereas the corrected rainfall become relatively large over the E–TP and during the monsoon season. The relative differences are almost more than 15% for all months and all subregions. Therefore, this correction is not negligible. Nevertheless, Fig. 7b shows that the magnitude of the new estimate for the entire plateau is only slightly greater than two previous estimates. This is probably related to the different periods and spatial averaging methods concerned in these studies, in addition to the precipitation correction in the present study.

c. Radiative convergence

As shown in Fig. 6c, RC in all months has negative values, which means that the so-called radiative convergence is actually radiative cooling. Surprisingly, the conventional and the new RC estimates are so different in both seasonality and magnitude. The RC estimate by the conventional method (using ISCCP–FD data) is generally smaller than the new estimate, except during several months over the W–TP. The former is less than the latter by 25% over the C–TP and the E–TP at annual scale. In an annual cycle, the former estimate nearly gives the minimum radiative convergence (or maximum radiative cooling) during the monsoon season [June–August (JJA)], when, by contrast, the new estimate gives the maximum radiative convergence (or minimum radiative cooling).

Figure 7c compares RC of the new estimate and two estimates in the literature for the entire plateau. Similar to the above conventional estimate, the previous estimates are generally significantly lower than the new one, particularly for summer or for the rainy season. These large differences highlight the primary importance to conscientiously evaluate the radiation data used for RC estimation. It is observed in Fig. 7c that the result in Zhao and Chen (2001) is close to the present one during winter. However, their difference in RC is as high as 20 W m$^{-2}$.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Region</th>
<th>Input</th>
<th>Period</th>
<th>Value (W m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ye and Gao (1979)</td>
<td>W–TP</td>
<td>CMA data</td>
<td>June</td>
<td>−220</td>
</tr>
<tr>
<td></td>
<td>E–TP</td>
<td>CMA data</td>
<td>May</td>
<td>−120</td>
</tr>
<tr>
<td></td>
<td>E–TP</td>
<td>Reanalysis data</td>
<td>May–June</td>
<td>−105</td>
</tr>
<tr>
<td>Wu and Zhang (1998)</td>
<td>Entire TP</td>
<td>Reanalysis data</td>
<td>Since May</td>
<td>(120–150)</td>
</tr>
<tr>
<td>Chen et al. (2003)</td>
<td>TP</td>
<td>CMA data</td>
<td>May–June</td>
<td>−120</td>
</tr>
</tbody>
</table>
during summer (JJA), when the former overestimated the surface net radiation.

d. Total heat source

Figure 6d shows there are fundamental differences in the total heat source estimated between the two methods. Regarding the seasonality, the conventional estimate shows the maximum TH occurs in May (over the W–TP) or June (over the C–TP and the E–TP), but the new estimate indicates that the maximum TH occurs in July for all subregions. The maximum TH in the conventional estimate is comparable to the new one over the W-TP, but it is significantly less than the latter over C–TP and E–TP. Figure 7d shows that the maximum value of the plateau-averaged TH occurs in May or June in three previous estimates, while it occurs in July in the new estimate. Zhao and Chen (2001) produced lower values of TH climatology, because of its lower RC and lower LH values. Of some interest is the fact that the climatology in Ye and Gao (1979), the earliest estimate to our knowledge, is closest to the new one. This is a coincidence, because its large positive biases in SH greatly cancel out its large negative biases in RC. The TH values in Yanai et al. (1992) are close to those in Zhao and Chen (2001) during JJA. Nevertheless, we note that there is an abrupt change of heat source from May to June of 1979 in Yanai et al. (1992), but this change is not seen in Zhao and Chen (2001) nor in other climatologic studies.

Two fundamentally different spatial patterns of TH in summer (JJA) were reported in the literature. One pattern is “TH decreases from the W–TP to the E–TP.” For example, Ye and Gao (1979) reported the value of TH climatology is 116 W m$^{-2}$ over the W–TP and 94 W m$^{-2}$ over the E–TP. In the summer of 1979, Yanai et al. (1992) gave a TH value of 78 W m$^{-2}$ over the W–TP and 71 W m$^{-2}$ over the E–TP. The other pattern is “TH increases from the W–TP to the E–TP,” which was concluded by studies using the data from First Qinghai-Xizang (Tibet) Plateau Meteorological Experiment (TIPMEX). For example, Chen et al. (1985) gave 46 W m$^{-2}$ of TH over the W–TP and 77 W m$^{-2}$ over the E–TP for the summer of 1979. Zhao and Chen (2001) produced TH climatology of 59 W m$^{-2}$ over the W–TP and 71 W m$^{-2}$ over the E–TP. In this study, we showed that the conventional method reproduced the former pattern (76 W m$^{-2}$ over the W–TP and 61 W m$^{-2}$ during JJA), whereas the new estimate yielded the latter pattern (82 W m$^{-2}$ over the W–TP and 111 W m$^{-2}$ over the E–TP during JJA).

A more detailed spatial distribution of TH during JJA is shown in Fig. 8. High TH (>100 W m$^{-2}$) is observed over the southern and eastern TP, while low TH (<100 W m$^{-2}$) over the northern and western TP. Another noteworthy result is that the newly estimated summertime TH values over the E–TP are higher than most previous estimates, mainly because of the much smaller radiative cooling in this study (Fig. 7c).

5. Heat source trend over the Tibetan Plateau

Figure 9 shows the linear trends in SH, RC, and TH for all subregions and all seasons, together with 95%
The trends in LH are not evident and thus omitted. The trends in the heat source terms over the W–TP are small and most of them exhibit large uncertainties and cannot pass the significance test ($p < 0.05$). By contrast, the trends over the C–TP and the E–TP are larger in magnitude and most of them pass the significance test. In general, both the conventional method and the new method produced negative trends in SH, RC, and TH; however, the new method gave weaker negative trends than the conventional one. Regarding the trends in SH, the new ones are about half to one-third of the conventional ones. The latter are believed overly estimated because of two causes. First, the bulk heat transfer coefficient has strong diurnal variations (small in the nighttime and large in the daytime) over the TP; thus, it is theoretically questionable to take the coefficient as a constant value throughout a day, as assumed in the conventional method and many previous studies. Second, because of climate change, the atmospheric instability actually increased over the TP because the wind speed steadily weakened and ground air temperature increased during the concerned period. Accordingly, the bulk heat transfer ($C_H$) may have increased with the enhancement of the atmospheric instability. This change would suppress the decreasing trend in SH, but this effect is not taken into account in the conventional method (Yang et al. 2011).

Based on Tibetan experimental data, Yang09 developed a new algorithm to calculate sensible heat flux from CMA data. This algorithm was applied by Yang et al. (2011) to estimate the SH trend over the TP. Figure 10 shows that the relative trends ($\%$ decade$^{-1}$) in the newly estimated SH are comparable to those given by Yang et al. (2011) but weaker than the conventional ones. As the new approach used herein and Yang09 are independent methods, their agreement in the SH trends indirectly supports each other.

Regarding the RC trends (Fig. 9b), they are dominated by large negative trends in the TOA radiation budget (Fig. 4a). The new RC trends are comparable to the conventional ones over the W–TP; however, the two become quite different over the E–TP, where the new ones are about half of the conventional ones. This difference is mainly caused by the use of different data sources of the surface net radiation. The former used LSM-simulated net radiation, which does not show a strong trend. The
conventional one used ISCCP–FD data, which yields a significant positive trend on the E–TP (not shown).

The TH trends are shown in Fig. 9c. The new TH trends are comparable to the conventional ones over the W–TP, but they become nearly half of the latter over the C–TP and the E–TP.

In summary, the conventional method may have notably overestimated the decreasing trends in the TP heat source terms.

6. Upscaling from stations to regional scale

The above estimates are based on surface data and satellite data at CMA stations. Given the TP’s vast area, the number of CMA stations is too small to be fully meaningful. Most of the stations were located on the C–TP and the E–TP, and there are only three stations on the W–TP. So, it is worth analyzing the spatial representativeness of the above estimate. In reality, upscaling the heat source terms from stations to the regional scale has not been addressed in the literature. In this study, a simple spatial upscaling with the aid of satellite products was given to evaluate the representativeness of the estimated climatology of the heat source terms and their trends. The spatial upscaling was applied to monthly-mean data, and the regional mean defined below is the value averaged over all grids with an elevation of more than 3000 m in each of the three subregions.

a. Upscaling approach

A hypothesis for the upscaling is that the spatial variability represented by relevant satellite products is reliable though their regional-mean values may be biased. This hypothesis seems realistic, as radiation and precipitation are mainly influenced by clouds, whose spatial distribution can be directly observed by satellites.

The regional mean of RC was upscaled with the aid of the GEWEX SRB data, as this satellite product has better accuracy than ISCCP–FD for the TP region (Fig. 3), as follows:

\[ RC_{reg} = RC_{sta} \frac{RC_{reg, srb}}{RC_{sta, srb}} \]

where \( RC_{reg} \) is the radiative convergence averaged over all grids in a region, \( RC_{sta} \) is the corresponding value averaged over all stations in the region (i.e., the estimate in section 3), \( RC_{reg, srb} \) is the RC value of GEWEX SRB averaged over all grids in the region, and \( RC_{sta, srb} \) is the corresponding value averaged over all stations in the region.

The spatial upscaling for LH is similar, as follows:

\[ LH_{reg} = LH_{sta} \frac{P_{reg, trmm}}{P_{sta, trmm}} \]

where \( LH_{reg} \) is the latent heat release averaged over a region, \( LH_{sta} \) is the value averaged over all stations in the region (i.e., the estimate in section 3), \( P_{sta, trmm} \) is the TRMM precipitation averaged over all stations in the region, and \( P_{reg, trmm} \) is the corresponding value averaged over all grids in the region.

SH is assumed to be proportional to surface net radiation at each month and each region. Its spatial upscaling is defined as follows:

\[ SH_{reg} = SH_{sta} \frac{Rn_{sfc, reg, srb}}{Rn_{sfc, sta, srb}} \]

where \( SH_{reg} \) is the SH value averaged over a region, \( SH_{sta} \) is the LSM-simulated SH value averaged over all stations in the region (i.e., the estimate in section 3), \( Rn_{sfc, reg, srb} \) is the surface net radiation from GEWEX SRB averaged...
over all grids in the region, and $R_{\text{nfc, sta_srb}}$ is the corresponding value averaged over all stations on the region.

\textit{b. Upscaling effects on the heat source estimate}

The GEWEX SRB radiation data and the TRMM precipitation data required in the upscaling are available for the period 1984–2004 and the period since 1998, respectively. So, the results for the common period (1998–2004) are presented below.

The heat source estimate after upscaling is shown in Table 3. Figure 11 compares the climatology of monthly-mean heat source terms before and after upscaling. The values after upscaling are generally less than before upscaling. This is related to the location of these stations. As seen in Fig. 1, there are more stations on the southern TP than on the northern TP; therefore, the estimated values before upscaling better represent the case in the southern TP. Because more precipitation and radiation take place on the southern TP than on the northern TP, the heat source is stronger over the southern TP, as can be seen in Fig. 8. As a result, the TH values after upscaling are mostly less than the ones before upscaling and should be more representative at the regional scale. For instance, their differences can be up to 20 W m$^{-2}$ in a region with sparse station distribution (e.g., the W–TP). This experience suggests that the uncertainty in the estimated heat source due to this scale effect may be comparable to that due to the methods for heat source calculation in section 3 and should be addressed. In addition, it is worth noting that the TH values decrease more over the W–TP, where CMA station distribution is much denser. Therefore, the regional contrast of TH between the W–TP and the E–TP is enhanced by the spatial upscaling.

On the other hand, the effects of the spatial upscaling seem rather insignificant on the trend estimation. As shown in Fig. 12, the interannual variability of JJA-mean and annual-mean TH values before upscaling is very similar to the counterpart after upscaling. This suggests that the trends in the heat source terms are generally not sensitive to the number of stations. In other words, the trends can be more reliably quantified from a limited number of stations than the climatology of the heat source.

\begin{table}[h]
\centering
\begin{tabular}{|c|c|c|c|c|c|c|c|c|c|c|c|c|c|c|c|}
\hline
Month & 1 & 2 & 3 & 4 & 5 & 6 & 7 & 8 & 9 & 10 & 11 & 12 & Annual \\
\hline
W–TP & 20 & 35 & 56 & 73 & 87 & 87 & 81 & 68 & 67 & 44 & 21 & 11 & 54 \\
C–TP & 21 & 37 & 58 & 73 & 77 & 66 & 66 & 52 & 43 & 36 & 31 & 23 & 17 & 44 \\
Avg & 18 & 32 & 51 & 66 & 71 & 64 & 56 & 46 & 42 & 30 & 19 & 12 & 42 \\
\hline
W–TP & 7 & 8 & 8 & 9 & 19 & 33 & 31 & 13 & 2 & 3 & 2 & 12 \\
C–TP & 4 & 5 & 12 & 32 & 66 & 84 & 85 & 44 & 14 & 2 & 2 & 30 \\
E–TP & 7 & 9 & 20 & 37 & 75 & 118 & 118 & 88 & 44 & 9 & 3 & 54 \\
Avg & 6 & 7 & 11 & 19 & 39 & 68 & 80 & 78 & 48 & 20 & 5 & 2 & 32 \\
\hline
W–TP & 88 & 76 & 61 & 42 & 29 & 36 & 36 & 51 & 70 & 69 & 80 & 86 & 60 \\
C–TP & 81 & 68 & 53 & 44 & 41 & 47 & 45 & 52 & 75 & 87 & 92 & 86 & 64 \\
E–TP & 82 & 68 & 55 & 43 & 42 & 49 & 50 & 59 & 76 & 87 & 92 & 87 & 66 \\
Avg & 84 & 71 & 56 & 43 & 37 & 44 & 44 & 54 & 74 & 81 & 88 & 86 & 63 \\
\hline
W–TP & 61 & 33 & 4 & 39 & 68 & 69 & 78 & 47 & 10 & 23 & 56 & 73 & 6 \\
C–TP & 57 & 27 & 10 & 40 & 68 & 85 & 91 & 76 & 6 & 42 & 66 & 68 & 10 \\
E–TP & 62 & 35 & 6 & 45 & 84 & 108 & 107 & 87 & 33 & 28 & 71 & 75 & 17 \\
Avg & 60 & 32 & 7 & 41 & 73 & 87 & 92 & 70 & 16 & 31 & 64 & 72 & 11 \\
\hline
\end{tabular}
\caption{Monthly-mean heat source terms (W m$^{-2}$) estimated in this study after spatial upscaling with the aid of GEWEX SRB radiation data and TRMM precipitation data, averaged over the period of 1998–2004.}
\end{table}

\section{Concluding remarks}

Atmospheric heating over the TP plays a profound role in the Asian monsoon system. Although many studies in the literature have focused on estimating the heating components and the total heat source, most of them lack substantial supports of observational data; therefore, existing estimates give a mixed picture, and some of them seem suspicious. In this study, recent advances in field data analyses, land model development, and satellite estimates of radiation and precipitation are utilized to produce a state-of-the-art estimate of the TP atmospheric heat source.

By comparing with the new estimate, we uncovered major biases in previous estimates. First, most previous studies overestimated the sensible heat flux in May and June. Some estimates are even higher than the observed surface net radiation, which violates the constraint of the surface energy balance at a monthly or annual scale. An
exception is the sensible heat flux presented in Zhao and Chen (2001), which is similar to the present one, as both estimates are constrained by the surface energy budget. Second, previous studies also greatly underestimated radiative convergence (or overestimated the radiative cooling) during the summer monsoon (or rainy) period. Nevertheless, the wintertime radiative convergence presented in Zhao and Chen (2001) is close to the present one. Third, the latent heat release was estimated based on CMA gauge-measured precipitation data in both the present study and many previous ones; therefore, their results...
are similar to each other. The present study considered a correction to account for the precipitation measurement loss and thus gives slightly higher latent heat release values than previous estimates.

Accordingly, the seasonality and regional pattern of the total heat source in the new estimate are different from previous studies. The total heat source in each subregion or the plateau reaches maximum during the mature stage of the monsoon season (July), whereas previous studies gave the maximum in the premonsoon season (May) or the beginning of the monsoon season (June), because of greatly overestimated radiative cooling in the monsoon season. The new estimate shows the maximum heat source occurred over the E–TP rather than over the W–TP, whereas most previous studies show an opposite regional pattern, because of the overestimated sensible heat flux on the W–TP before the onset of the monsoon. A spatial upscaling with the aid of satellite data further enhances the spatial pattern of the heat source distribution given by the new estimate.

This study also investigated the trends in the heat source terms. A recent estimate by DW08 showed strong negative trends in sensible heat and radiative convergence as well as the total heat source. This study confirms the negative trends, but the magnitudes in their work seem greatly overestimated. Specifically, their sensible heat trend is 2–3 times the new one averaged over the plateau, because atmospheric stability has been weakening in recent decades but their estimate omitted its effect on the heat exchange (Yang et al. 2011). The trend in radiative convergence given by DW08 is similar to the new one over the W–TP, but it becomes twice the new one over the C–TP and the E–TP, because of biases of the surface net radiation used in their study. As a result, the trend in the total heat source given in their study is about $-12 \text{ W m}^{-2} \text{ decade}^{-1}$ for the entire plateau, while the value given herein is $-7 \text{ W m}^{-2} \text{ decade}^{-1}$.

We believe that this observation-supported study may reduce uncertainties in the heat source estimate over the TP, which may contribute to relevant model evaluations, processes analyses, and climate change studies. Nevertheless, further experiments and studies are encouraged to reach consensus on the estimated heating components and the total heat source. Particularly, we recommend more investigations on the radiative forcing, as its accuracy significantly affects the heat source estimation.

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