THE ATMOSPHERIC HEAT BUDGET IN SUMMER OVER ASIA MONSOON AREA

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ABSTRACT

For better understanding the mechanism of monsoon formation and designing the numerical simulation of the general atmospheric circulation, a new approach of calculating atmospheric radiation is proposed to investigate the distribution of the atmospheric heat source, and the budget of heat component is recalculated. The results show that there is a tremendous atmospheric heat source region over central India, northeast of the Bay of Bengal, east of the South China Sea and about 10 °N at the west Pacific, among which the heating center with a maximum heating rate of 8 °C/day is located over the Bay of Bengal and the average rate in the Plateau is about 1 °C/day.

I. INTRODUCTION

It is well known that the main heat source driving the global atmosphere is over the tropic, particularly the Asia monsoon area in summer. Yeh et al. (1957)[1] first proposed that there is an atmospheric heat source over the Tibetan Plateau in summer. Later Chen et al. (1965)[2] pointed out that in summer the atmospheric heating center with maximum heating rate 2.6°C / day is over the southeast part of the Tibetan Plateau and the west part of Yunnan Province, and that the heating rate over the central part of the Plateau is only about 1°C/day. Katayama (1967)[3] studied the budget of radiative energy. Using some empirical formulas, Kubota (1970)[4] calculated the budget of the atmospheric radiation. The distribution of the atmospheric heat source given accordingly shows that the atmospheric heating center is located over Southeast Asia. The formulas employed in his computation of radiation are rather simple, and the data are insufficient, so the results are preliminary. Based on a diagnostic equation of heating rate, Li et al. (1983)[5] calculated the atmospheric heating rate from the stream function and the temperature. Their results indicated that the heating center in the Asian summer monsoon area with a heating rate larger then 3°C/day is over the northern Bay of Bengal and the central Plateau with a heating rate of about 1.4°C/day is in the margin of the heat source region.

For better understanding the mechanism of monsoon formation and designing the numerical simulation of the general atmospheric circulation, further research has been done in this paper.

II. METHOD OF CALCULATION AND DATA

The atmospheric heat source (HS) consists of four atmospheric heat components: the absorption of solar radiation ΔS, the atmospheric long-wave radiation budget ΔF (the difference between the surface effective radiation ER and the long-wave radiation F_{00} at the top of the atmosphere), the surface eddy flux of sensible heat SH and the latent heat LP released by condensation in the process of precipitation. This can be expressed as

$$HS = \Delta S + \Delta F + SH + LP,$$

where $HS>0$ (<0) is the atmospheric heat source (sink). The manipulation of ΔS and ΔF is somewhat tedious, a summary description would be given later.

1. *Data Treatment*

The data are taken at five levels in the atmosphere and at the earth's surface. The vertical coordinate σ is used, where $\sigma = p/p_s$, p is the air pressure and p_s the surface air pressure. The five levels considered are 0.1, 0.3, 0.5, 0.7 and 0.9, respectively. The observed data at eight standard levels (1000, 850, 700, 500, 300, 200, 150, 100 hPa) are interpolated into the five σ levels.

The optical mass of water vapor is corrected by a integral factor $(p/p_0)^{0.72}$ with $p_0=$ 1013.5 hPa. For CO_2, the integral factor of correction of the optical mass takes the form $(p/p_0)^{1.00}$. The concentration of CO_2 is taken to be 3.14×10^{-4}.

In the entire atmosphere, only two layers of clouds are considered: one is between $\sigma=0.8$ and 0.9 and the other is between $\sigma=0.4$ and 0.6. In respect of long-wave radiation the clouds are treated as blackbody radiators. Assuming that C is the total cloud cover and C_1 and C_2 are the fractional cloud cover at middle and lower level in the sky respectively, C_1 could be obtained from C and C_2, which are available in the surface observtion. It could be expressed as

$$C_1 = \frac{C-C_2}{1-C_2} \quad .$$

According to the two layers of clouds overlapping or not, there are four regions in the sky: a) the clear region, with fractional area $(1-C_1)(1-C_2)$; b) the region where only middle level clouds exist, with fractional area $C_1(1-C_2)$; c) the region where only lower level clouds exist, with fractional area $C_1(1-C_1)$; and d) the region where both middle and lower level clouds exist, with fractional area C_1C_2. For a certain level, the insolation or radiative absorption is taken as the weighted average of the insolation of the four regions by the fractional areas. Denote the net long-wave radiations at this level of these four regions by F_{00}, F_{c_10}, F_{0c_2}, $F_{c_1c_2}$ respectively. Assuming no interactions between the four regions, the overall net radiation F could be written as:

$$F=(1-C_1)(1-C_2)F_{00}+C_1(1-C_2)F_{c_1O}+C_2(1-C_1)F_{Oc_2}+C_1C_2F_{c_1c_2}$$

The absorption of solar radiation could be similarly obtained.

2. *The Atmospheric Absorption of Solar Radiation*

The incident solar radiation at the top of atmosphere is divided into two parts, a scattered part SS and an absorbed part SA:

$$SA=0.349S, \qquad \lambda>0.9\mu,$$
$$SS=0.651S, \qquad \lambda\leqslant 0.9\mu.$$

The scattered part could only be absorbed by the ground surface. The total absorption at the surface ΔSS_{10} is given by

$$\Delta SS_{10}=SS_{10}\left[\frac{(1-C_1)(1-C_2)(1-\alpha_A)(1-\alpha_G)}{1-\alpha_A\alpha_G}\right.$$
$$+\frac{C_1C_2(1-R_1)(1-R_2)(1-\alpha_A)}{1-(R_1R_2+R_1\alpha_G+R_2\alpha_G-2R_1R_2\alpha_G)}+\frac{C_1(1-C_2)(1-R_1)(1-\alpha_G)}{1-R_1\alpha_G}$$
$$\left.+\frac{C_2(1-C_1)(1-R_2)(1-\alpha_G)}{1-R_2\alpha_G}\right],$$

where $\alpha_A=0.085-0.25074\log_{10}\left(\frac{p_s}{1023.5}\sec Z\right)$ is the albedo of the atmosphere due to Rayleigh scattering, $\alpha_G=0.15$ is the surface albedo to the scattering radiation, $R_1=0.54$ and $R_2=0.66$ are the albedos of middle level and lower level cloud layers to the scattering, respectively.

The atmospheric absorption of solar radiation is the sum of direct and indirect absorptions. The absorption between $\sigma=0.6$ and 0.8 generally is A_1+A_2. For clear sky A_1 is the absorption of the direct solar radiation and A_2 the absorption of the radiation reflected by the earth's surface. For the lower level A_2 is the absorption of radiation reflected by the top of clouds. For two separated cloud layers, A_1 is the absorption of direct radiation across the upper atmosphere and middle level cloud layer and A_2 is the absorption of the radiation reflected by the lower level cloud layer. For other cases, A_1 and A_2 could be assigned accordingly.

In computation, the albedos of middle and lower cloud layers are 0.46 and 0.5 respectively. The absorption of cloud layer is obtained by using absorption formula of water vapor. The optical mass of water vapor in the cloud layer is 10 (for middle level) or 8 (lower level) times of that under condition of clear sky. For the atmosphere under the cloud layer, its optical mass is multiplied by a factor of 1.66 which is an argumentative factor for diffuse radiation below the cloud. The absorption formula of water vapor derived by Chen et al.[6] is used in this paper.

The principal absorption of solar radiation by CO_2 occurs in the upper troposphere. In this study, the absorption of reflected radiation by CO_2 is neglected and the computation is under the condition of clear sky.

The mean incident insolation at the outer limit of the atmosphere is

$$S_0=0.349\frac{K}{D^2}\overline{\cos Z}\,S_{00},$$

where S_{00} is the solar constant (1.94 cal/cm^2 min^{-1}); D the ratio of the mean earth-sun distance to the actual distance and for July it takes the value 1.01644; K the ratio of the daytime to the length of a whole day (24 hr); Z the zenith angle and bar denotes the mean in the period of daytime. Let T be the number of days starting at 1 January (for 16 July, $T=197$), φ the latitude, δ the declination angle and ω_0 the hour angle at sunset, we have

$$\delta=23.5\sin 2\pi\left(\frac{T-82}{365}\right),$$

$$K=\frac{\omega_0}{2\pi}=\frac{1}{2\pi}\cos^{-1}(-\tan\varphi\tan\delta),$$

$$\overline{\cos Z}=\frac{1}{\omega_0}\int_{-\omega_0}^{\omega_0}\cos Z d\omega=\sin\varphi\sin\delta+(\cos\varphi\cos\delta\sin\omega_0)/\omega_0.$$

3. *Calculation of the Atmospheric Long-Wave Radiation*

According to Manabe et al.[7] the net radiation F_σ passing through σ level can be expressed as

$$F_\sigma=\pi B_S-\pi B_C\varepsilon\{u^*(T_{00})-u^*(T_\sigma),T_c\}\pi dB-\int_{B_C}^{B_{00}}\tilde{\varepsilon}\{u^*(T_{00})-u^*(T_\sigma),T\}\pi dB$$
$$-\int_{B_{00}}^{B_S}\tilde{\varepsilon}\{u^*(T_{00})-u^*(T_\sigma),T\}\pi dB,$$

where B represents blackbody radiation, subscriptions S and C refer to the earth's surface and a certain level in the air, respectively. Besides,

$$B=81.25042956\times10^{-2}T^4\mathrm{cal/(cm^2\ min)};$$

u^* is the optical mass corrected by pressure; $\tilde{\varepsilon}$ the mean emissivity; ε the mean emissivity above level C. For water vapor, we have

$$\tilde{\varepsilon}_f=\frac{1}{\pi\frac{dB}{dT}}\int_0^\infty\pi\frac{dB_\nu}{dT}\{1-\tau_f(l_\nu u)\}d\nu,$$

$$\varepsilon_f(u^*,220\ K)=\frac{1}{\pi B}\int_0^\infty\pi B_\nu\{1-\tau_f(l_\nu u)\}d\nu.$$

The integration could be numerically calculated by using the trapezoid formula. The net radiation at a certain level is obtained by weighted averaging the radiations at the above-mentioned four regions in the sky on the consumption that the top and the cloud layer are treated as blackbody and the weights are the fractional areas of these four regions. This approach is the same as that for the computation of the absorption of the short-wave radiation.

Around the H_2O–CO_2 overlapping wavelength 15μ, emissivity is corrected by $\Delta\varepsilon$[8], and ozone is neglected in our computation. The emissivity and overlapping correction for H_2O and CO_2 are estimated according to Sasamori[9].

5. *Computation of LP and SH*

LP is converted from monthly mean precipitation. Sensible heat flux *SH* could be obtained from

$$SH=C_p\rho_0C_0|V|(T_s-T_a),$$

where C_p is the specific heat of air at constant pressure, ρ_0 the air density near the ground, $|V|$ the velocity of surface wind, T_s the temperature of ground soil, T_a the air temperature at surface and C_D the drag coefficient. Different values of C_D are assigned by many authors. In this study, we take C_D to be 8.0×10^{-3} for the Plateau with elevation higher than 3000 m, 5.0×10^{-3} for the area with elevation between 1000 and 3000 m, 3×10^{-3} for the lower elevation less than 1000 m, 1.5×10^{-3} for ocean. These values of C_D are employed in our computation of *SH* for China and the adjacent ocean area. For other regions, the *SH* distribution given by Schutz et al. (1971)[10] is directly referred.

The data used in the computation are based on the observations of 77 stations in China and 84 stations in other countries. Mean values of radiosonde data of foreign stations (1961—1974) taken from Boogaard (1979)[11], are used and the mean values of the stations in China for the same period are computed from observations. The precipitations for all stations are from climatological data supplied by State Meteorological Administration of China and taken from Monsoon Experiment, GARP Publ. No. 18[12] for the foreign stations. The climatology of precipitations for longer period in the Philippines and Northeast Burma is used. The cloudiness data are also taken from Surface Station Climatology Atlas published by the State Meteorological Administration of China.

III. THE ATMOSPHERIC HEAT BUDGET

Fig. 1 shows the computed atmospheric long-wave radiation budget ΔF in terms of temperature change rate. From Fig. 1 it is clearly seen that over the Tibetan Plateau there is a cooling center with the heating rate less then -3.0 °C/day, and the heating rates for Lhasa and Tingri are -3.2 °C/day and -3.1 °C/day respectively. Southern Tibet, Yunnan Province and west Guizhou are within the area with the heating rate less than -2.5 °C/day. The heating rates in continental east China are about -2°C/day. The results are quite agreeable to that obtained by Chen et al. (1965)[2]. The cooling center of the whole Asia is located over the Bay of Bengal and the South China Sea with cooling rate less than 2.0 °C/day. The cooling rate caused by long-wave radiation of CO_2 is about 0.2—0.3°C/day.

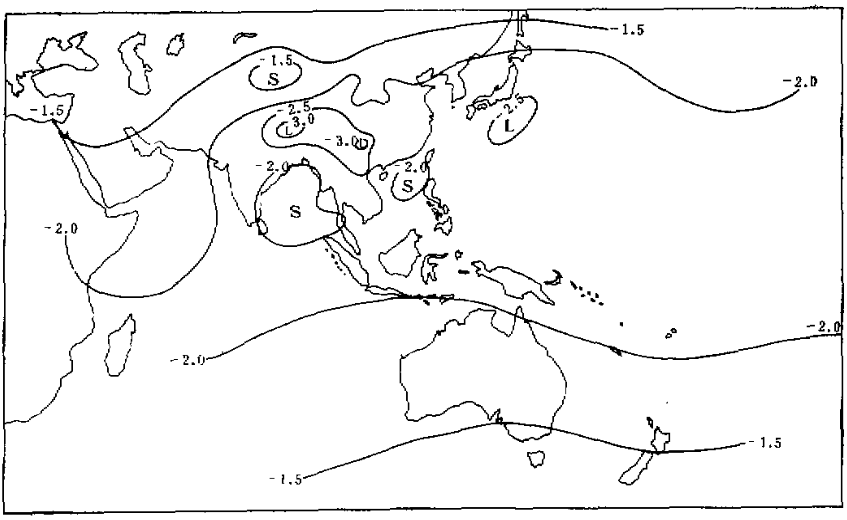

Fig. 1. The mean cooling rate in July due to atmospheric long-wave radiation (unit: °C/day).

The absorptions of solar radiation by the atmosphere ΔS in terms of heating rate are indicated in Fig. 2. It is shown that there is a lowest value over the Tibetan Plateau. The heating rates are 0.74°C/day and 0.64°C/day for Lhasa and northwest of the Plateau respectively. This is evidently associated with low moisture over the Plateau. On the contrary, the highest values (larger than 1°C/day) occur over India, the Bay of Bengal and the South China Sea. Thus, the heating rates converted from the difference of overall atmospheric radiations are less than -2.0°C/day over the Plateau and about 1°C/day over the Bay of Bengal, the South China Sea and West Pacific tropics. The calculated values for the area over the Plateau are close to that obtained by Chen et al[2].

Shown in Fig. 3 are the heating rates due to the release of latent heat in the process of precipitation. It is indicated that the higher values present over the northern part of the west coast of India, the east of the Bay of Bengal, the east of the South China Sea, West Pacific ITCZ and from the area between the Changjiang and Huaihe Valleys up to southern Japan, among which a maximum value about 10°C/day occurs over the east of the Bay of Bengal. There might be a heating center over the northeast of the Bay of Bengal, the northwest of Burma, Bangladesh and the northeast of India. The mean precipitation over the northwest coast of Burma in July is about 1000 mm, which could make temperature rise to 8.2 °C. The heating rates are about 0.2—0.5°C/day and 1.0—1.5°C/day for the west part and northeast part, of the Tibetan Plateau respectively, which are only one sixth of that over the Bay of Bengal due to the little rain fall. This is understandable if one notices that the monthly precipitation in Lhasa in July is about 70—100 mm and the thickness of the whole atmosphere on the Plateau is about 600 hPa while the rain fall in the northeast coast of the Bay of Bengal is 1000 mm and the thickness of the atmosphere is about 1000 hPa. A heating center of 8°C/day lies over the east of the South China Sea, and is the secondary maximum center in the area of the monsoon activity. An area with low heating rates situated at about 15 °N in India, the western part of the Bay of Bengal and the west of the South China Sea coincides with the area with small cloudiness in the satellite cloud picture, meanwhile the high heating rates occur in the area of strong precipitation.

The sensible heat fluxes from the surface are depicted in Fig. 4 (solid lines). High values could be found over the Arabian Peninsula, the Iran Plateau and the Tibetan Plateau. Over the central and western parts of the Tibetan Plateau occurs the highest value about 250 cal cm^{-2} day^{-1}, which is larger than that given by Gao et al. (1981)[13]. It can be clearly seen that in the Northern Hemisphere south of 22.5°N the averaged value of SH is about 10—20 cal cm^{-2} day^{-1}, which is much less than that over the Plateau, and SH almost vanishes over the Arabian Sea. However, the result for the Tibetan Plateau is somewhat larger than that we would expect. This might be caused by an overestimated parameter $C_D=8\times10^{-3}$. As a matter of fact, although plateaus are high in elevation and generally have rougher surface than plains or oceans, the ground on plateaus is rather flat. A value of 4—5×10^{-3} for C_D might be appropriate. If so, the actual sensible heat flux would be smaller than that indicated in Fig. 4. The heating rates converted from the computed SH are shown by the dash lines in Fig. 4, in which it can be seen that the temperature would rise by 1.0—1.5 °C per day over the Tibetan Plateau and by 1.7°C per day at Nagqu, the highest for the Plateau. In the Tibetan Plateau area, the surface sensible heat flux is the most important term in all the heat budget components of the atmospheric source, and comparable to the latent heat in the central and eastern parts but much larger than the latter in the western part.

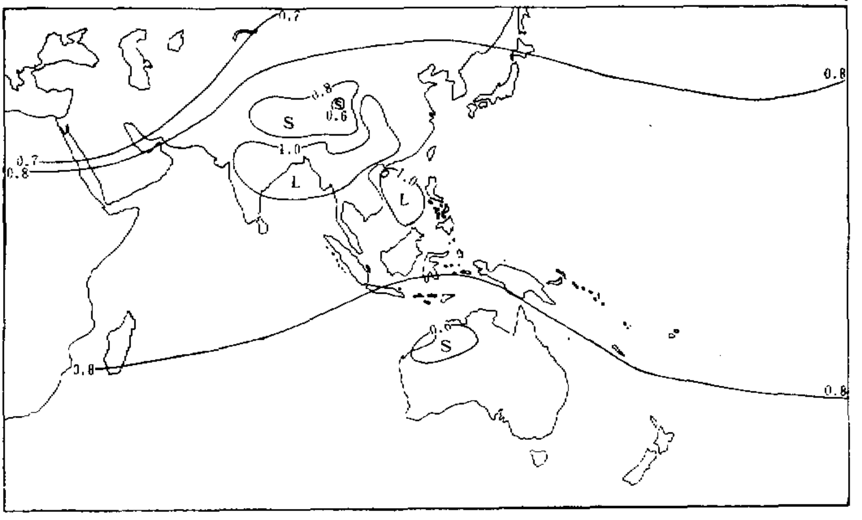

Fig. 2. The mean heating rate in July due to the atmospheric absorption of solar radiation (unit: °C/day).

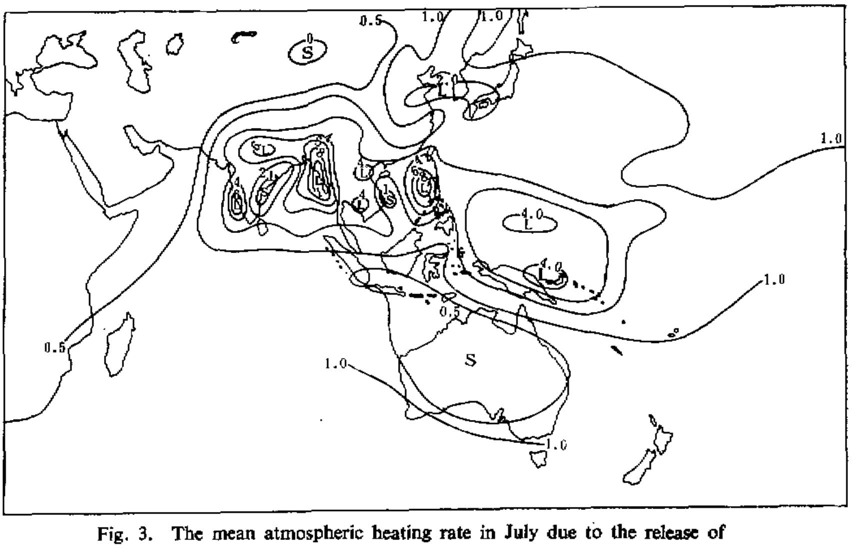

Fig. 3. The mean atmospheric heating rate in July due to the release of latent heat by rain fall (unit: °C/day).

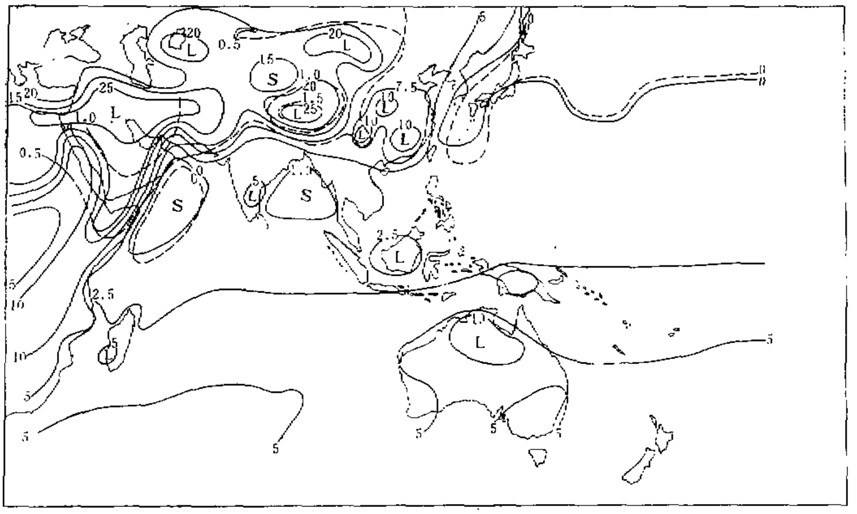

Fig. 4. The mean atmospheric heating rate in July due to eddy sensible heat flux at surface (solid lines in unit: 10 cal cm^{-2} day^{-1}; dash lines in unit: °C/day).

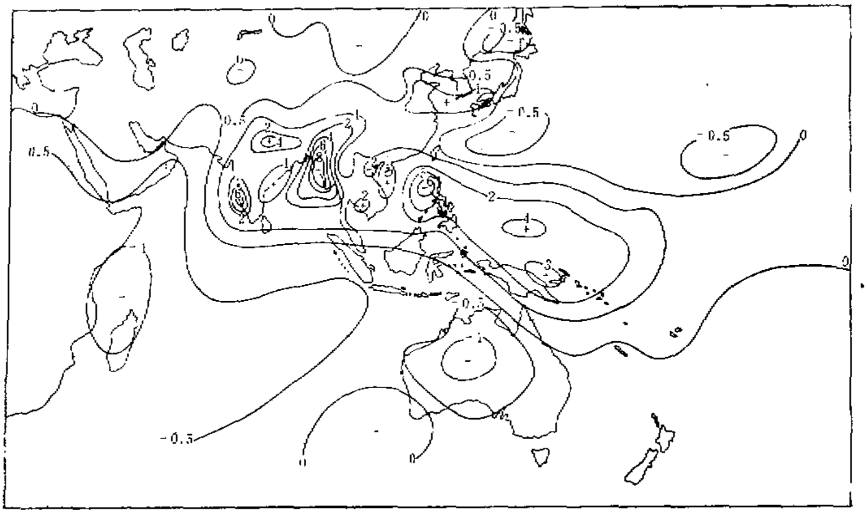

Fig. 5. The distribution of the mean atmospheric heat source for July (unit: °C/day, − denotes heat sink, + denotes heat source).

IV. THE DISTRIBUTION OF THE ATMOSPHERIC HEAT SOURCE

Fig. 5 gives the distribution of the atmospheric heat source for July averaged over several years in terms of heating rate(solid lines). The main atmospheric heat sources are all situated in the areas of monsoon troughs in southern Asia and West Pacific ITCZ. In the west coast of India, monsoon trough in Indian sub-continent, eastern Bay of Bengal (including northwest coast of Burma, northeast of India), eastern South China Sea and south of west Pacific ITCZ, strong heat sources could be found. In east coast of India (western part of the Bay of Bengal) and western South China sea the weaker heat sources are located. Among these, the ones with the heat rates of 8°C/day and 6°C/day over eastern Bay of Bengal and eastern South China Sea are the strongest. The two heat sources have very important effects on the structure of the atmospheric circulations in Asia. Krishnamurti (1971)[14] indicated that over the Bay of Bengal occupies a high center of velocity potential with strong divergence in the upper troposphere. So, it is a divergent heat source. Murakami (1978)[15] showed that in summer of 1970—72 the velocity potential center at 200 hPa level is over Philippines. These two velocity potential centers overlap the mean atmospheric heat sources. The heat sources have definite effects in maintaining the velocity potential. Meanwhile the divergences in the upper troposphere transport the thermal energy, produced by the atmospheric heat sources to the sourrounding areas, thus eventually maintaining the heat sources. The maximum heating rates in the two sources could reach as high as 8 °C/day and 6 °C/day respectively. For the heat source over the Bay of Bengal, the monthly mean precipitation of 1000 mm in July in the Northwest Burma reported by Yeh et al. (1980) [16] corresponds to a heat rate of 9 °C/day. By taking into account the radiative cooling rate 1 °C/day, the temperature could actually rise by 8°C per day. For the same consideration, the heating rate due to the latent heat release over the north Philippines, which is embedded in the heat source centers of the South China Sea, is estimated to be 8°C/day and the potential temperature rise would be 7°C per day.

Although the Tibetan Plateau could be considered as an atmospheric heat source, it contributes to the temperature rise much less compared with the two heat sources mentioned above, in the south of the Plateau only 0.7—1.0°C per day as indicated in Fig. 5 and no more than 1.0—1.3 °C /day if only long-wave radiation of water vapor is taken into account as in early literatures. For the whole Plateau the average heating rate of the atmospheric heat sources is about 0.6 °C/day (0.9 °C/day if the effect of long-wave radiation of CO_2 is ruled out). Over the western part of the Titetan Plateau sensible heat flux is the principal component of heat sources while over the east, sensible heat flux nearly equals the latent heat release (about 1.5 °C/day for each). The total amount of the thermal energy due to these two heat components is just cancelled by the cooling effect of long-wave radiation. Thus, the overall atmospheric heating approaches the heating due to the absorption of solar radiation. This explains why the heat source over the Plateau is not so strong. In July it has less precipitation over the Plateau, only 75—100 mm even in the southern part compared with 1000 mm in the Bay of Bengal. The rate of temperature rise only due to the latent heat is about 5—8 times as much as that due to the Plateau (noticing that the thickness of the atmosphere over the Plateau is only 600 hPa, the heating rate over the Plateau should be 1.6 times of that over plains under the same heating circumstance). In addition, because of lacking moisture in the Plateau atmosphere it absorbs very little solar radiation, while it is cooling down due to the long-wave radiation. This makes it easy to understand that there is a huge difference of heating rates between the Plateau and the Bay of Bengal by a factor of 8. Therefore, the Tibetan Plateau should not be regarded as a strong atmospheric heat source. Instead, the principal heat sources are located over the Bay of Bengal and the South China Sea. As a matter of fact, many striking weather phenomena are observed over these two areas.

The above arguments are based on the precipitation data of a few scarcely distributed stations. In better and denser observations will occur the high precipitations, which would affect the evaluation of the term of latent heat, the major term in these components of heat source. In this case, our statements about the atmospheric heat source over the Tibetan Plateau should be adjusted correspondently.

By examining the pattern of the atmospheric heat sources in Fig. 5, one can see that the heat source over the Plateau is only the northward extended part of the center of the atmospheric heat source over the Bay of Bengal. In fact, monsoon activities greatly affect the fluctuation in the intensity of the atmospheric heat source over the Plateau, the latter could be intensified only in case monsoon intrudes up the Plateau region and causes rain fall.

The atmospheric cooling centers are mainly located in the fallowing areas: 1) the western part of the Arabian Sea; 2) the Indian Ocean west of 120°E and south of 4°N; 3) the southern Pacific south of 20°S and Australia; 4) the Pacific north of 20°N (In the coast of China, heat sink might invade around 25°N, and the continent at the same latitude is a weak source, apparently related to the subtropic high); and 5) the Asian continent north of 45 °N.

In Fig. 5 is also shown a weak source belt from Huanghe-Huaihe Valley up to South Japan with a maximum heating rate about 1.6°C/day which obviously corresponds to the rainy belt in July.

V. CONCLUSIONS

From the above discussion, we can draw the following conclusions:

(1) In summer the atmospheric heat source centers are situated in the area of tropic monsoon trough and ITCZ. In Asian monsoon area they are Indian sub-continent, the northeast of the Bay of Bengal up to northeast India, th east of South China Sea and west Pacific tropic ITCZ. Among those, the strongest one is over the northeast of the Bay of Bengal with a heating rate as high as 7—8 °C/day.

(2) In summer there is an atmospheric heat source over the Tibetan Plateau and it could not be regarded as a major source for atmospheric heating, because the heating rate is no more than 1 °C/day with an average 0.5—0.6 °C/day.

(3) The Arabian Sea, west Pacific subtropic high area, west Indian Ocean between 3—30°S and Australia are the atmospheric heat sinks.

In recent years, the role of the Tibetan Plateau as an atmospheric heat source has been highly addressed by many authors. Of course, it is not to be neglected. In our view, there is no enclosed center of heat source over the Tibetan Plateau and it is only a margin of the principal heat source embedded in monsoon area. Thus, in numerical experiments and simulations the heat sources, not only over the plateau but also over the monsoon area, should be set, otherwise the results would be unmeaningful. A single source over the plateau would overestimate the thermodynamic effect of the plateau. The numerical experiment[17] showed that if the heating center was placed at 20°N south of the plateau, the results obtained would be agreeable to the observations.

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