Atmospheric Diabatic Heating and Summertime Circulation in Asia–Africa Area[①]

Li Weiping (李伟平)

State Laboratory of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)
Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100029

Theo Chidiezie Chineke

Department of Physics, Imo State University, PMB 2000, Owerri, Nigeria

Liu Xin (刘 新) and Wu Guoxiong (吴国雄)

State Laboratory of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)
Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing 100029

(Received January 19, 2000; revised May 29, 2000)

ABSTRACT

Utilizing data from NCEP / NCAR reanalysis, the summertime atmospheric diabatic heating due to different physical processes is investigated over the Sahara desert, the Tibetan Plateau, and the Bay of Bengal. Atmospheric circulation systems in summer over these three areas are also studied. Thermal adaptation theory is employed to explain the relationship between the circulation and the atmospheric diabatic heating.

Over the Sahara desert, heating resulting from the surface sensible heat flux dominates the near–surface layer, while radiative cooling is dominant upward from the boundary layer. There is positive vorticity in the shallow boundary layer and negative vorticity in the middle and upper troposphere. Downward motion prevails over the Sahara desert, except in the shallow near–surface layer where weak ascent exists in summer. Over the Tibetan Plateau, strong vertical diffusion resulting from intense surface sensible heat flux to the overlying atmosphere contributes most to the boundary layer heating, condensation associated with large–scale ascent is another contributor to the lower layer heating. Latent heat release accompanying deep convection is critical in offsetting longwave radiative cooling in the middle and upper troposphere. The overall diabatic heating is positive in the whole troposphere in summer, with the most intense heating located in the boundary layer. Convergence and positive vorticity occur in the shallow near–surface layer and divergence and negative vorticity exist deeply in the middle and upper troposphere. Accordingly, upward motion prevails over the Plateau in summer, with the most intense rising occurring near the ground surface. Over the Bay of Bengal, summertime latent heat release associated with deep convection exceeds longwave radiative cooling, resulting in intense heating in almost the whole troposphere. The strongest heating over the Bay of Bengal is located around 400 hPa, resulting in the most intense rising occurring between 300 hPa and 400 hPa, and producing positive vorticity in the lower troposphere and negative vorticity in the upper troposphere. It is also shown that the divergent circulation is from a heat source region to a sink region in the upper troposphere and vice versa in lower layers.

Key words: Atmospheric diabatic heating, Summer, Circulation

1. Introduction

It has long been recognized that thermal contrast between land and sea is the main

①This work was jointly supported by "National Key Program for Developing Basic Sciences" G1998040904 by NSFC projects 49805003, 49635170, 49823002, and 49825504.

driving force of monsoon (Krishnamurti and Ramanathan, 1982; Chen and Dell'Osso, 1986), which emphasizes the significance of the external forcing. If we take the earth and the atmosphere as a whole into account, both the boundary forcing by the sensible heat flux to the overlying atmosphere and the latent heat release in the air column contribute to the thermal–driving atmospheric circulation. As the highest landform in Asia, the Tibetan Plateau exerts significant thermal influence on general circulation (e.g., Yeh et al., 1957; Chen et al., 1985a, b; Yanai et al., 1992; Yang et al., 1992; Ye and Wu, 1998). Recently, Wu (2000) and his colleagues established the " thermal adaptation" theory to elucidate the maintaining mechanism of general circulation over the Tibetan Plateau. Their main concern was on the non–uniformity of diabatic heating in the vertical and its climate effect (e.g., Wu and Liu, 2000; Liu et al., 2000; Wu et al., 2000). Actually, the sensitivity of atmospheric response to the vertical variation of heating has been shown in model results for both tropical (Hartmann et al., 1984) and extratropical (Trenberth, 1983) latitudes. In this paper, the vertical distribution of atmospheric diabatic heating in Asia–Africa area and their relationship with the general circulation are studied. The situations of three critical regions: the Tibetan Plateau (hereafter referred to as TP), the Sahara desert and the Bay of Bengal (BOB in short in the following) are the main concern.

In next section, the data used and the method of analysis are presented. The seasonal variations of the atmospheric diabatic heating and the components of the heating in July over the three representative areas are investigated in Section 3. The possible relationship between the diabatic heating and atmospheric motion is discussed in Section 4. Section 5 is devoted to the study of the divergent circulation in the Asia–Africa area, which is followed by a summary and discussion in Section 6.

2. Data and method

The data employed in this study consist of 18–year (1980–1997) monthly mean diabatic heating rates (referred to as Q in the following), relative vorticity (ζ), horizontal wind (u,v), and vertical p–velocity (ω) from NCEP / NCAR reanalysis. The Q data are in σ–coordinate scheme, and the rest are in p–coordinate. In order to facilitate analysis, the diabatic heating rates are interpolated to p–coordinate assuming its linear variation with log (p) in the vertical. The heating rates are the output of NCEP assimilation system, which are calculated from the temperature tendency resulting from various physical processes such as longwave and shortwave radiation, deep convection, large scale condensation, shallow convection, and vertical diffusion. For the sake of brevity, positive and negative Q is referred to as heating and cooling, respectively. Time average is taken for 18 years to get monthly mean of each variable. Regional means are calculated over the Sahara desert (18–30°N, 10°W–30°E), the Tibetan Plateau (28–37°N, 80–100°E) and the Bay of Bengal (15–25°N, 80–98°E), representing dry, elevated and rainy areas, respectively.

3. Atmospheric diabatic heating

Figure 1 displays the seasonal vertical variation of regional mean diabatic heating averaged over the three areas. For the Sahara desert (Fig. 1a), there is heating all the year round below the level of about 820 hPa, with the most intense heating occurring at about 900 hPa from early March through late May. Above 700 hPa, persistent cooling dominates. Although the strongest near–surface heating in the year does not appear in July, positive Q reaches its

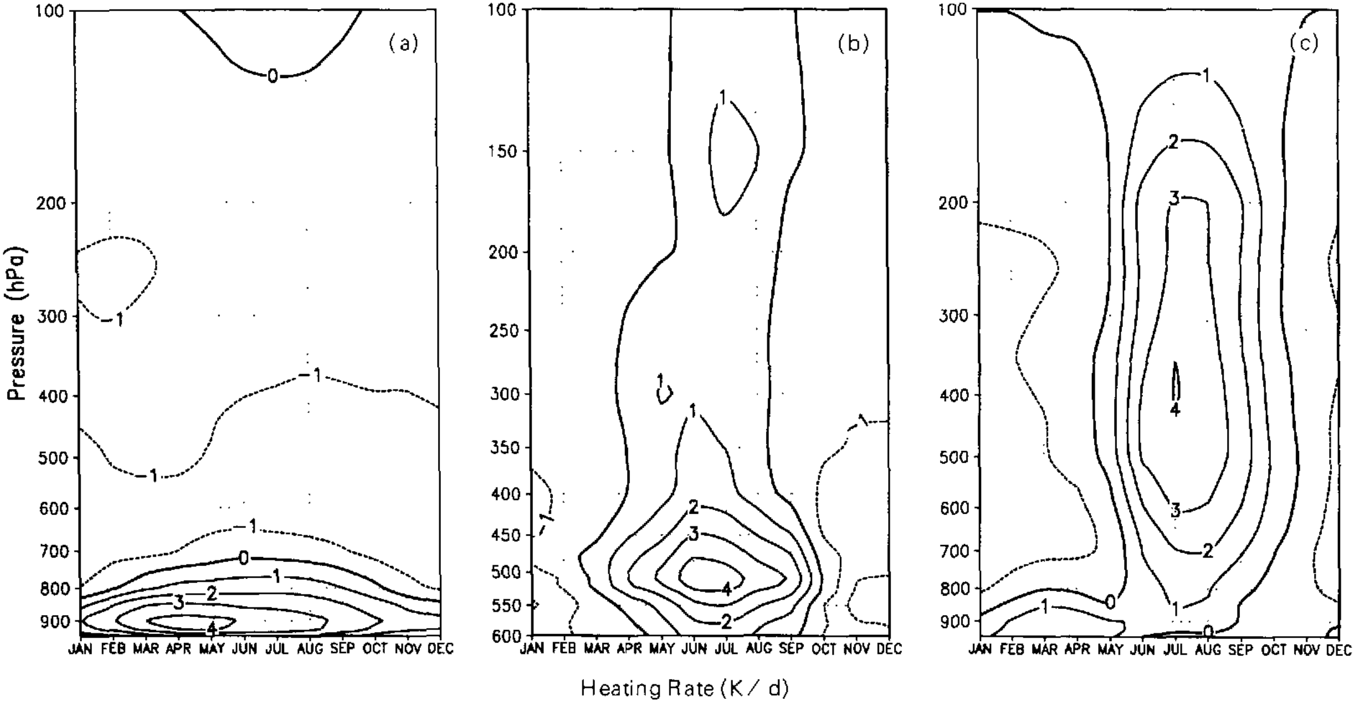

Fig. 1. Time–height section of the 18–year (1980–1997) monthly mean diabatic heating rates averaged over three regions. (a) the Sahara desert (10°W–30°E, 18–30°N, the left panel); (b) the Tibetan Plateau (80–100°E, 28–37°N, the central panel) and (c) the Bay of Bengal (80–92°E, 15–25°N, the right panel). The interval of isopleths is 1 K d^{-1}.

maximum in the vertical (about 720 hPa) during this period.

Over the TP (Fig. 1b), one can identify the whole tropospheric heating from early May through early August during the summer rainy season, which agrees well with the results of Yao et al. (1982). The intense heating exists in the near–surface layer with the maximum at about 500 hPa, in which layer the air experiences the longest period of heating from mid–February through September. The most significant difference from that of the Sahara desert is the middle and upper level heating in summer versus cooling over the latter location.

The air of the BOB above 600 hPa experiences intense heating from early May through mid–October (Fig. 1c), while the rest of the year see cooling. The situation below the level of 900 hPa is almost contrary to that above 600 hPa, the winter through spring heating there can be attributed to the surface sensible heat flux from the relative warm ocean to the cool continental origin air flow. Different from the above two areas, the strongest heating over the BOB occurs at the upper troposphere around 400 hPa, suggesting the significance of condensation heating associated with deep convection, which is clearly shown in Fig. 2c.

In order to investigate various contributions to the overall diabatic heating, ingredients due to different physical processes in July are displayed in Fig. 2. Over the dry Sahara desert area (Fig. 2a), vertical diffusion due to the strong surface sensible heating dominates below the level of 800 hPa, while above this level longwave radiative cooling is more pronounced. Solar radiation is another ingredient contributing to the overall heating in the whole troposphere, in spite of its relatively small quantity, whereas heating due to convection and condensation can be ignored. The overall feature indicated by the vertical heating profile is that $Q > 0$ in the lower layer below 700 hPa, while $Q < 0$ in the middle and upper troposphere, with the maximum existing around 900 hPa, which is in good agreement with the result of Yanai and Tomita (1998). Key reasons for the overall cooling in the middle and upper air column are from the high reflectivity of the surface and thermal emissions for the generally cloud–free surface over the dry Sahara desert area.

Over the Plateau (Fig. 2b), situations are rather complex. None of the ingredients can be ignored except shallow convection. It is apparent that longwave radiation is the main cooling term in the troposphere. The heating due to shortwave radiation is relatively small. The latent heat release associated with deep convection contributes most of heating in the upper layers from 350 hPa to 150 hPa. The heating in lower layers is mainly attributed to large–scale condensation from 500 hPa to 400 hPa, and to vertical diffusion below 500 hPa. The strong heating (about 3.8K d^{-1}) due to vertical diffusion near the ground surface implies the significance of the surface sensible heat flux over the TP. The total heating is positive in the whole troposphere over the Plateau, which is consistent with the traditional viewpoint that the Tibetan Plateau is a heat source in summer (Yeh et al., 1957; Chen et al., 1985b; Yanai et al., 1992). The overall heating of the atmosphere over the TP reaches its maximum slightly below 500 hPa, decreases gradually upward to the upper troposphere till 200 hPa, which agrees well with that of Yanai and Tomita (1998). The second maximum at 150 hPa is mainly due to deep convection.

For the Bay of Bengal (Fig. 2c), the critical factor for diabatic heating from 800 hPa upward is deep convection, which reaches its maximum at about 400 hPa, thus produces the most intense overall heating (about 4K d^{-1}) in this layer. Both the longwave and shortwave radiation appear nearly constant in the vertical for the deep cloud cover over there. Like the case over the TP area, the general feature of diabatic heating is $Q > 0$ in almost the whole

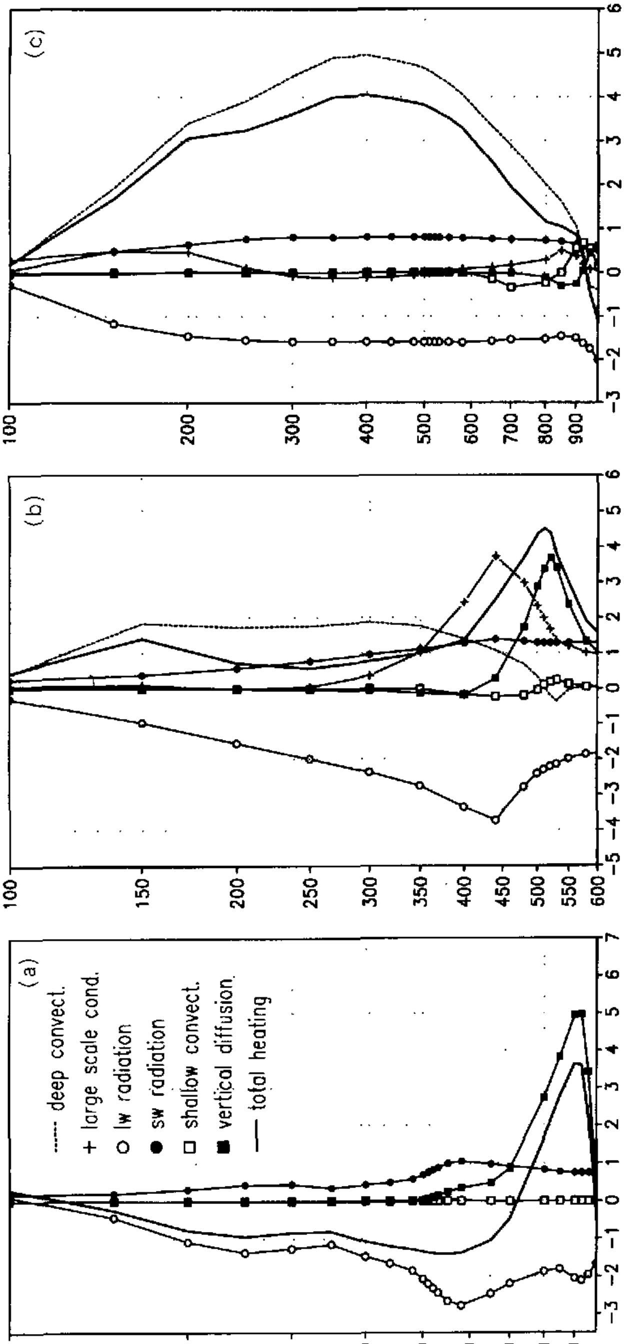

Fig. 2. Vertical profile of regional averaged heating ingredients due to different physical processes in July (1980–1997 mean) for the same three areas as in Fig. 1. Unit is K d^{-1}. Legends are displayed in the left panel.

troposphere, except in the near–surface layer where radiative cooling prevails, suggesting that the BOB is one of the most intense atmospheric heat source regions in the Northern summer. This has long been recognized by previous researchers (e.g., Schaack et al., 1990; Yanai et al., 1992, 1998).

From the above analysis, we can infer that the TP and the BOB are heat source regions in summer, whereas the Sahara desert is a sink region of heat.

4. Relationship between atmospheric diabatic heating and atmospheric motion

Atmospheric response to the vertical variation of heating has been suggested to be sensitive from model results (Trenberth, 1983; Hartmann et al., 1984). According to the complete form of vorticity tendency equation developed by Wu and Liu (1999), the following approximate relationship is satisfied between the tendency of vertical component of relative vorticity and the vertical distribution of heating rate in the atmosphere:

$$\partial\zeta / \partial t \propto \partial Q / \partial z \ , \tag{1}$$

where ζ represents the vertical component of relative vorticity. It can be inferred that $\partial Q / \partial z > 0$ is below the level of maximum heating, while $\partial Q / \partial z < 0$ is true above that level. From (1), it can be followed that $\partial\zeta / \partial t > 0$($\partial\zeta / \partial t < 0$) is satisfied below (above) the maximum heating level. In other words, there is positive vorticity generation in lower layers and negative vorticity generation in upper layers relative to the level of maximum heating. This is substantiated by the results shown in Fig. 3. The shallow near–surface layer of positive vorticity over the Sahara desert and the TP, and rather deep layer of positive vorticity over the BOB from surface to about 400 hPa are consistent with $\partial Q / \partial z > 0$ in the corresponding layers shown in Fig. 2. On the contrary, the deep upper layer of negative vorticity upward from 800 hPa over the Sahara, from 500 hPa over the TP, and from 300 hPa over the BOB coincides with the situation of $\partial Q / \partial z < 0$ in the corresponding layers displayed in Fig. 2. These verify the theoretical analysis and agree well with the numerical simulations of Wu and Liu (2000).

From the thermodynamic equation of atmospheric circulation, on the time scale of monthly mean, the tendency term can be ignored. If the relatively small advection term in the subtropical area is omitted, the vertical velocity and diabatic heating satisfy the following approximate relationship (Ding, 1991):

$$(\gamma_d - \gamma)w \approx Q / C_p \ , \tag{2}$$

where γ_d is dry adiabatic lapse rate, γ is environmental temperature lapse rate, w represents the vertical velocity, and C_p is specific heat at constant pressure. Over the Sahara desert, the surface sensible heating affects only the thin planetary boundary layer. Above the middle level of the troposphere, radiative cooling is dominant. The low–level heating reaches its highest position at about 700 hPa (Fig. 2a) in July. From that layer upward, $Q < 0$ is satisfied. The atmospheric stratification is usually stable, i.e. $\gamma_d - \gamma > 0$, following from (2) that $w < 0$. It implies that the radiative cooling of subtropical atmosphere is balanced by the adiabatic heating associated with sinking (Fig. 3a). This is one of the important mechanisms that maintain the North African anticyclone in addition to contributing to the dynamics of the Asian–African monsoon. On the contrary, the boundary layer rising is due to the shallow near–surface heating.

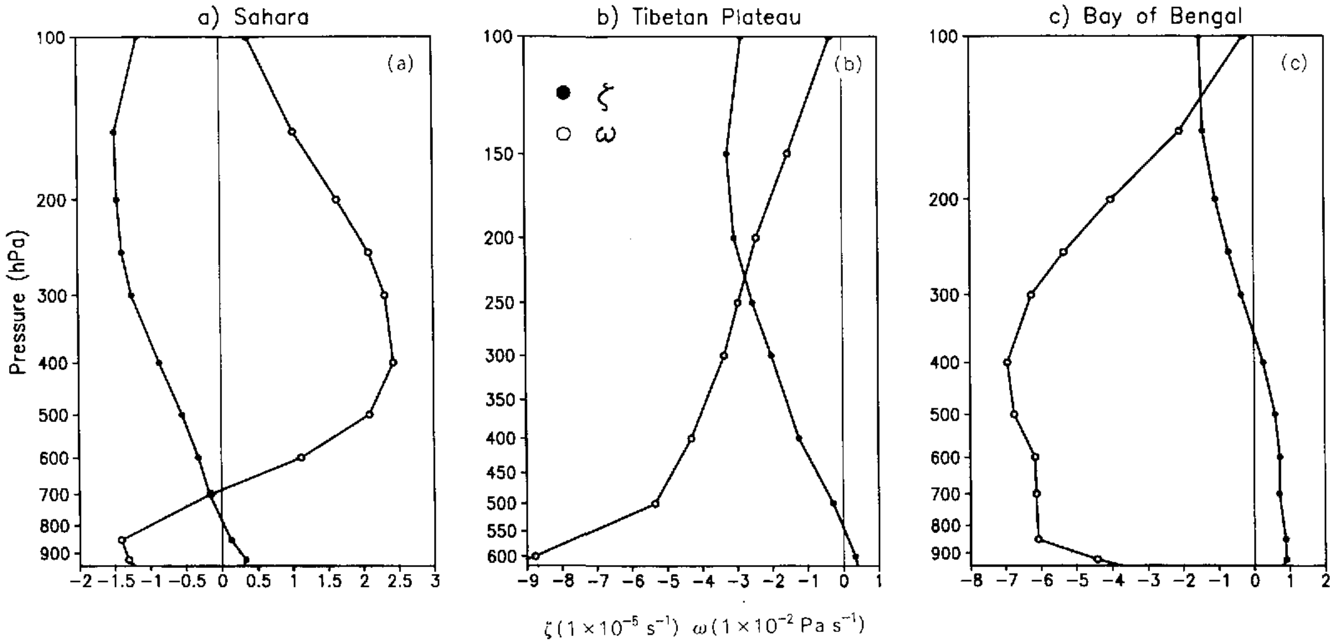

Fig. 3. Vertical profile of regional averaged relative vorticity (solid circle, Unit in 10^{-5} s^{-1}) and vertical p-velocity (open circle, Unit in 10^{-2} Pa s^{-1}) in July (1980–1997 mean) for the same three areas as in Fig. 1.

The theory of "two–stage" thermal adaptation proposed by Wu et al. (2000) is employed to explain the rising over the TP and BOB. At the first stage, the surface sensible heating produces cyclonic circulation and ascent in the near–surface layer, which is applicable to both areas. The strong surface heating over the TP produces intense low level ascent (Fig. 3b). In the second stage, with plenty supply of moisture, large amount of latent heat release associated with cloud formation and precipitation generates intense rising, especially for the BOB (Fig. 3c). The strongest upward motion and the highest reach of the maximum rising over the BOB among the three areas clearly elucidate the fully developed second–stage thermal adaptation over there.

5. Divergent circulation

It has been suggested that divergent air flow is from atmospheric heat source region to the sink region in upper layers and vice versa in lower layers (Johnson et al., 1985). Regional air motion is discussed in this section with the divergent circulation being our main concern. From the geographic distribution of vertical component of relative vorticity and divergent wind (Fig. 4), one can easily identify the upper level (200 hPa, Fig. 4a) strong negative vorticity center anchored over the TP, suggesting the powerful South Asian anticyclone in summer. The divergent center over the BOB is in significant contrast with the weak convergence toward the northeastern Sahara desert (Fig. 4a). The divergent flow linking these two regions can be easily identified. The negative vorticity over the Sahara desert, the TP and the BOB are consistent with that shown in Fig. 3.

At 600 hPa level which is approximately the surface layer for the TP, a positive vorticity center and air convergence can be identified over the Plateau (Fig. 4b). Whereas the negative vorticity over the western and southern Sahara desert and the positive vorticity over the BOB are in sharp contrast, which is consistent with the two different vertical profiles of diabatic heating shown in Fig. 2a and Fig. 2c.

At 925 hPa, positive vorticity and air convergence to the southern Sahara and to the BOB are apparent (Fig. 4c), exhibiting the boundary layer positive vorticity shown in Fig. 3a and Fig. 3c. Comparing Fig. 4b with Fig. 4c, one can infer that the convergence to the southern Sahara is confined to a rather shallow layer. On the contrary, the convergence to the BOB and the cyclonic circulation over there exist in a quite deep layer, which is in agreement with the vertical profile of diabatic heating (Fig. 2a, c). The deep convection over the BOB supplies plenty of moisture to feed the heavy rain over there in summer, which results in the fully developed second–stage of thermal adaptation (Wu et al., 2000). The divergent flow from heat sink region–the northern Sahara and the eastern Mediterranean Sea as well as the Arabian Sea to heat source regions–the southern flank of Sahara, the northern India and the BOB can be identified.

Figure 5 displays the vertical divergent circulation over Asia–Africa area. The rising motion occurring equatorward of the TP is at once apparent. The ascent flow diverges both equatorward and poleward above 300 hPa at about 15°N, where the intense summertime heat source is located. Part of the rising flow over the TP moves poleward, the other part sinks between 40°N and 48°N below 300 hPa, contributing to the local dry climate. The southward branch of the divergent flow traverses the equator above 300 hPa, sinking in the region of heat sink (not shown) in the Southern Hemisphere (Fig. 5a). If we check the vertical–zonal circulation (Fig. 5b), the ascent flow over northern India and the BOB turns westward above 300 hPa, sinks in the broad longitudinal band spanning from the eastern Sahara, Saudi

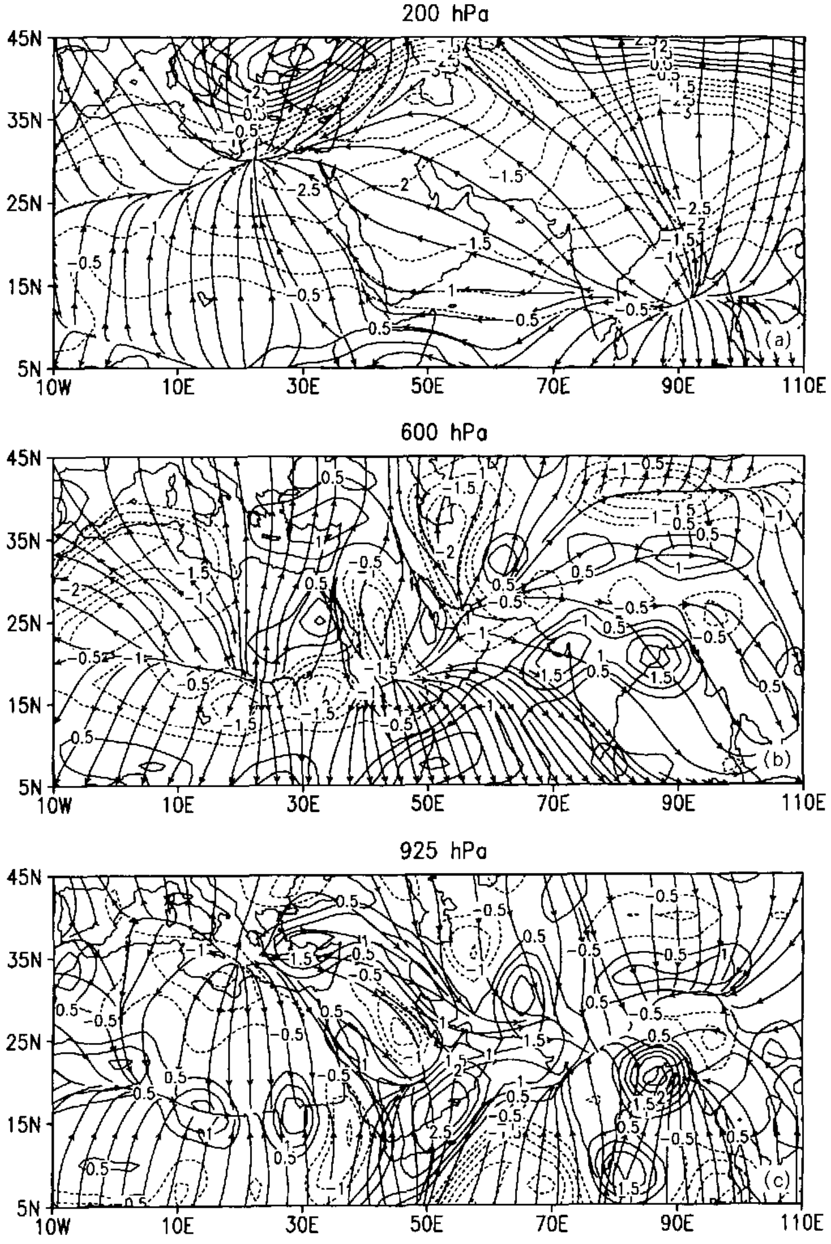

Fig. 4. Geographical distribution of 18-year (1980–1997) July mean relative vorticity in contours and divergent wind in streamline over Asia–Africa area at (a) 200 hPa, (b) 600 hPa, and (c) 925 hPa. Solid and dashed lines represent positive and negative values, respectively. Unit is 10^{-5} s^{-1} for relative vorticity. Contour interval is 0.5, and the contour zero is omitted.

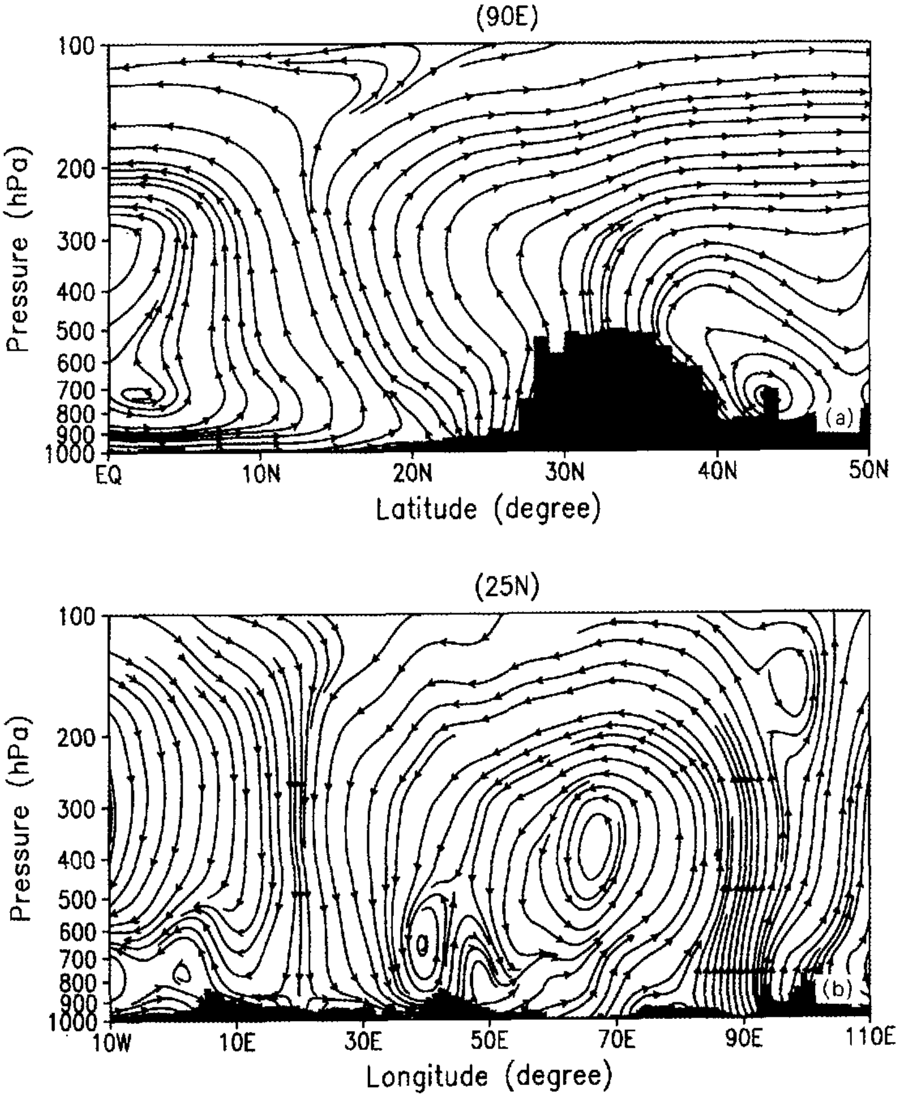

Fig. 5. Vertical cross-section along (a) 90°E and (b) 25°N for 18-year (1980–1997) July monthly mean divergent circulation in streamline.

Arabia to the northern Arabian Sea. The other branch of sinking motion over the Sahara originates from the Atlantic Ocean (not fully shown). The returning flow below 700 hPa from the eastern Sahara through the Arabian Sea to India and the BOB is also illustrated.

From Fig. 4 and Fig. 5, we can infer that the divergent circulation is from the heat source regions, i.e., the TP and the BOB to heat sink region (the Sahara desert) at upper level, and the returning flow at lower level is in the reverse direction.

6. Summary and discussion

Based upon the observational study of atmospheric diabatic heating and summertime circulation in Asia–Africa area, the following conclusions can be drawn:

1. Over the Sahara desert, heating resulting from the surface sensible heat flux dominates the near–surface layer, while radiative cooling is dominant above the boundary layer. The overall diabatic heating is positive in the shallow near–surface layer, but negative in the middle and upper troposphere. The strongest heating occurs at the layer immediately above the ground, accordingly, there is negative vorticity in almost the whole troposphere except the boundary layer. To balance the atmospheric cooling, downward motion prevails over the Sahara desert, except in the shallow near–surface layer, where there exists weak ascent in summer.

2. Over the TP, large surface sensible heat flux to the overlying atmosphere results in strong vertical diffusion, which contributes most to the intense near–surface heating. Condensation associated with large–scale ascent is another contributor to the lower layer heating. Latent heat release associated with deep convection is the chief heating ingredient in the middle and upper troposphere, whereas longwave radiative cooling is the critical counteragent in the whole troposphere under consideration. The overall diabatic heating is positive in the whole troposphere in summer, with the most intense heating located at the boundary layer. The convergence and positive vorticity in the shallow near–surface layer and divergence and negative vorticity in the deep middle and upper troposophere are consistent with the theoretical analysis of Wu et al. (2000), and the weather experience as well. Accordingly, upward motion prevails over the TP in summer, with the most intense rising occurring near the surface.

3. Over the BOB, summertime latent heat release accompanying deep convection dominates over longwave radiative cooling, resulting in intense heating in almost the whole troposphere. The strongest heating over the BOB is located around 400 hPa, resulting in the most intense rising between 300 hPa and 400 hPa, and positive vorticity in the upper troposphere and negative vorticity in the lower troposphere.

4. The horizontal and vertical divergent circulation shows the close connection between atmospheric heat source and sink regions. The divergent flow is from heat source to sink region at upper level, whereas at lower level the flow is reversed from sink to source region.

It should be pointed out that the data used in this study are the output of NCEP assimilation system, the reliability of heating rates which depends highly on the model is open to question over data sparse area such as the Tibetan Plateau. But the general features of the diabatic heating in the three selected locations are in good agreement with previous studies using different data sets. Furthermore, the vertical profile of diabatic heating over the TP is consistent with the circulation distribution, and is in accordance with the theoretical analysis of Wu et al. (2000). It thus provides a lot of information about the nature of the atmospheric heating by quantitatively studying the various components due to different physical processes.

What we investigated in this work is the climate mean situation, the interannual variation of atmospheric diabatic heating is rather significant as has been reported by Schacck and Johnson (1994) and Yanai and Tomita (1998). The interannual variability of diabatic heating components and the associated variation in atmospheric circulation is an interesting topic for further study.

REFERENCES

Chen L., T. Duan, and W. Li, 1985a: Heat source variation and characteristics of energy budget over the Tibetan Plateau during the 1979 summer. *Acta Meteorologica Sinica*, **43**, 1−11.

Chen L., E. R. Reiter, and Z. Feng, 1985b: The atmospheric heat source over the Tibetan Plateau: May−August 1979. *Mon. Wea. Rev.*, **113**, 1771−1790.

Chen, S. J., and L. Dell'Osso, 1986: Numerical experiments on the sensitivity of the monsoon circulation to differential heating. *Quart. J. Roy. Meteor. Soc.*, **112**, 93−112.

Ding, Y., 1991: *Advanced Meteorology*, China Meteorological Press, Beijing, 792 pp (in Chinese).

Hartmann, D. L., H. H. Hendon, and R. A. Houze, 1984: Some implications of the mesoscale circulations in tropical cloud clusters for large−scale dynamics and climate. *J. Atmos. Sci.*, **41**, 113−121.

Krishinamurti, T. N., and Y. Ramanathan, 1982: Sensitivity of the monsoon onset to differential heating. *J. Atmos. Sci.*, **39**, 1290−1306.

Liu Xin, Wu Guoxiong, Li Weiping, and Liu Yimin, 2000: The summertime heating over the Tibetan Plateau and the maintenance mechanism of large−scale circulation. *Progresses in Natural Science* (to be published).

Johnson D., R. Townsend, and M. Wei, 1985: The thermally coupled response of the planetary scale circulation to the global distribution of heat sources and sinks. *Tellus*, **37A**, 106−125.

Schaack, T., D. Johnson, and M. Wei, 1990: The three−dimensional distribution of atmospheric heating during the GWE. *Tellus*, **42A**, 305−327.

Schaack, T., and D. Johnson, 1994: January and July global distributions of atmospheric heating for 1986, 1987 and 1988. *J. Climate*, **7**, 1270−1285.

Trenberth, K. E., 1983: Interactions between orographically and thermally forced planetary waves. *J. Atmos. Sci.*, **40**, 1126−1152.

Wu Guoxiong, and Liu Yimin, 2000: Thermal adaptation, overshooting, dispersion, and subtropical anticyclone. I: Thermal adaptation and overshooting. *Chinese J. Atmos. Sci.*, **24(4)**, 433−446 (in Chinese).

Wu Guoxiong, Liu Ping, Liu Yimin, and Li Weiping, 2000: Impacts of the sea−surface temperature anomly in Indian Ocean on the subtropical anticyclone. "two−stage" thermal adaptation in the atmosphere. *Acta Meteorologica Sinica*, **58(5)**, 513−521 (in Chinese).

Wu Guoxiong, and Liu Huanzhu, 1999: Complete form of vertical vorticity tendency equation and slantwise vorticity development. *Acta Meteorologica Sinica.*, **57**, 1−15 (in Chinese).

Yanai, M., C. Li, and Z. Song, 1992: Seasonal heating of the Tibetan Plateau and its effects on the evolution of the Asian summer monsoon. *J. Meteor. Soc. Japan*, **70**, 319−351.

Yanai, M., and T. Tomita, 1998: Seasonal and international variability of atmospheric heat sources and moisture sinks as determined from NCEP−NCAR reanalysis. *J. Climate*, **11**, 463−482.

Yang Weiyu, Ye Duzheng, and Wu Guoxiong, 1992: The influence of the Tibetan Plateau on the thermal and circulation fields over East Asia in summer. II: Main features of the local and large−scale vertical circulation fields. *Chinese J. Atmos. Sci.*, **16**, 287−301.

Yao, L. C., Luo, S. W., Wang A. Y., 1982: The characteristics of the atmospheric heating field distribution and its annual variation in East Asia. *Plateau Meteorology*, **1(3)**, 1−11 (in Chinese).

Yeh, T. C., Luo, S. W., Chu, P. C., 1957: On the heat balance and circulation structure in the troposphere over the Tibetan Plateau and its vicinity. *Acta Meteorologica Sinica.*, **28**, 108−121 (in Chinese).

Ye Duzheng, and Wu Guoxiong, 1998: The role of the heat source of the Tibetan Plateau in the general circulation. *Meteor. Atmos. Phys.*, **67**, 181−198.

亚—非季风区非绝热加热与夏季环流关系的诊断研究

李伟平 Theo Chidiezie Chineke 刘 新 吴国雄

摘 要

基于热力适应理论，本文利用 NCEP／NCAR 再分析资料对撒哈拉沙漠、青藏高原和孟加拉湾地区的非绝热加热与夏季环流进行了诊断研究。在非洲撒哈拉沙漠地区，以感热输送为主的加热仅局限于近地面层，边界层以上的大气则以辐射冷却占优势。因而除了边界层内存在着浅薄的正涡度和微弱的上升运动以外，整个对流层几乎都维持负涡度并盛行下沉运动。对于青藏高原地区，强大的表面感热通量引起的垂直扩散是近地面大气加热的主要分量，与大尺度上升运动相关的凝结潜热对低层大气的加热也有一定的贡献。长波辐射造成的对流层中、上层大气的冷却则主要由深对流潜热释放来补偿。夏季高原地区总非绝热加热是正值，且最大加热率出现在边界层内。低空大气辐合产生正涡度，而中、高层大气辐散伴有较强的负涡度。因而高原盛行上升运动，最大上升运动位于近地面层。夏季孟加拉湾地区的深对流凝结潜热释放远大于长波辐的冷却作用，因而整个对流层几乎都保持较强的非绝热加热。400hPa 层附近的最大加热率引起 300–400hPa 最强的上升运动。对流层上层是负涡度区，而中、低层为正涡度区。结果还表明，垂直和水平辐散环流与大气的热源和热汇区密切相联：在高层，辐散气流从热源区流向热汇区；在低层则相反，气流从热汇区流向热源区。

关键词：大气非绝热加热，夏季，环流