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NUMERICAL EXPERIMENTS OF THE INFLUENCES OF ANOMALY HEATING OVER THE TIBETAN PLATEAU ON THE FORMATION AND EVOLUTION OF THE ASIAN SUMMER MONSOON

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A three-level global atmospheric general circulation model (AGCM) has been used to study the impacts of the anomaly heating over the Tibetan Plateau on the formation and evolution of the Asian summer monsoon. The control experiment reproduces the most features of the observation in the Asian summer monsoon region. Two numerical experiments are performed with the anomaly snow cover over the Plateau in winter and without atmospheric diabatic heating over the Plateau in summer, respectively. The results show that a larger snow albedo than the normal on the Plateau in winter will keep the ground colder and cause the circulation anomaly. Its effects can expend to the whole summer and make the reduction of precipitation in the middle and lower reaches of Changjiang (Yangtze) River while it enhances in the South China Sea. When we remove the atmospheric diabatic heating over the Plateau in summer, the Indian summer monsoon will disappear and the east Asian summer monsoon will be weakened Therefore, it is revealed that the responses of the summer monsoon circulation and its evolution to the abnormal heating over the Plateau are remarkable.

Key Words: Numerical Experiments; Anomaly Heating; Tibetan Plateau; Asian Summer Monsoon

Introduction

Tibetan (Qinghai-Xizang) Plateau is the largest and highest plateau with the most complex geographical features in the world. The position of the Plateau locates over the zone of westerlies in winter and over the transitional zone of easterlies and westerlies in summer. The earth-atmosphere system near the Plateau is a cold heat source in winter and a heat source in summer, the differences of the circulation patterns over the Plateau and its neighbouring area features of synoptic processes in Asia may be attributed to its existance. Thus the Tibetan Plateau has special dynamic and thermodynamic effects on the global and regional atmospheric motions in different seasons.

There have been a lot of studies on the Tibetan Platéau and its influences on the atmospheric circulation[1-4]. Especially during May to August of FGGE (1979), the Qinghai-Xizang Plateau Meteorological Experiments (QXPMEX) was made by China and a lot of meteorological data were obtained. By using these data, Chinese meteorologists have done a lot of work and our knowledge about the Plateau meteorology has been improved obviously. However, there are many problems still with us.

In this paper, we use a three-level global atmospheric general circulation model (AGCM)[5] to study the anomaly heating effects of the Tibetan Plateau on the formàtion and evolution of the Asian summer monsoon. Based on this model, three experiments have been conducted. Comparing the results obtained from the experiments, the effects of the heat forcing of the Plateau are discussed.

Brief Discription of the Model and Experimental Schemes

The model we use is a three-level global atmospheric general circulation model including a planetary boundary layer (PBL). The thickness of the PBL is 50 hPa and from the top of PBL to 200 hPa is free atmosphere. The free atmosphere is divided into two levels. For the PBL and the free atmosphere level we use the independent σ-coordinate system respectively (σ' for the PBL and σ for the free atmosphere level):

$$\sigma = \frac{P - P_T}{P^*}$$

and

$$\sigma' = \frac{P - P_s + 50}{50},$$

where P_T is the constant pressure at top of the model (P_T= 200 hPa), P_s is the pressure at the earth's surface and P^* is the thickness of the free atmosphere

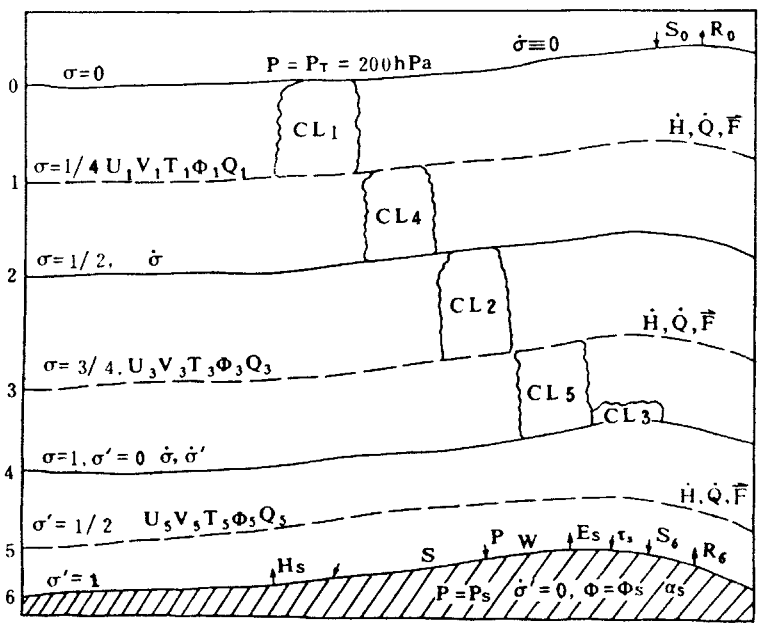

Fig 1 The vertical structure, principal variables and 5 cloud types of the three-level AGCM

level ($P^* = P_s - P_T - 50$), 50 hPa is the thickness of the PBL. The vertical structure, the basic variables and cloud types of the model are shown in Fig. 1, where u, v, T, q, Φ are the principal variables, P_T the pressure at the model top, P_s the pressure at the earth's surface, S and R the net downward solar and net upward terrestrial radiation at the top of the model atmosphere (subscript 0) or at the earth's surface (subscript 6), as the surface albedo, H, Q and $\mathbf{F}$ the diabatic heating, moisture source and friction, H_s the surface sensible heat flux, P the precipitation rate, E_s the surface moisture flux, τ_s the surface momentum flux, S the snow mass, and w the ground wetness. CL1-CL5 are 5 basic cloud types, 0-4 is free atmosphere level and 4-6 is PBL.

Differential Equations

For free atmosphere we use the σ-coordinates and the horizontal momentum equation in vector form and thermodynamic energy equation can be written in the form:

$$\frac{\partial}{\partial t}(P^*\mathbf{V}) + \nabla(P^*\mathbf{V}\mathbf{V}) + \frac{\partial}{\partial \sigma}(P^*\mathbf{V}\sigma) + f\mathbf{K} \times P^*\mathbf{V} + P^*\nabla\Phi + P^*\sigma\alpha\nabla P^* = P^*\mathbf{F} \quad \ldots (1)$$

and

$$\frac{\partial}{\partial t}(P^*C_pT) + \nabla\cdot(P^*C_pT\mathbf{V}) + \left(\frac{P}{P_{00}}\right) \times \frac{\partial}{\partial \sigma}(P^*C_p\sigma v) - P^*\sigma\alpha\left(\frac{\partial P^*}{\partial t} + \mathbf{V}\cdot\nabla P^*\right)$$

$$= P^*H, \quad \ldots (2)$$

where

$$v = Y\left(\frac{P_{00}}{P}\right)^x, \quad x = R/C_p = 0.286.$$

The moisture continuity equation is

$$\frac{\partial}{\partial t}(P^*q) + \nabla\cdot(P^*q\mathbf{V}) + \frac{\partial}{\partial \sigma}(P^*q\dot{\sigma}) = P^*\dot{Q} \quad \ldots (3)$$

The state equation is

$$\alpha = \frac{RT}{P} \quad \ldots (4)$$

The hydrostatic balance equation is

$$\frac{\partial}{\partial \sigma}(\boldsymbol{\Phi})+P^*\alpha=0 \qquad \ldots (5)$$

and the mass continuity equation is

$$\frac{\partial}{\partial t}(P^*)+\nabla\cdot(P^*\mathbf{V})+\frac{\partial}{\partial \sigma}(P^*/\dot{\sigma})=0 \qquad \ldots (6)$$

For the PBL we use σ' coordinate in the constant thickness level, the horizontal momentum equation can be written as follows:

$$\frac{\partial}{\partial t}(\mathbf{V})+\nabla(\mathbf{V}\mathbf{V})+\frac{\partial}{\partial \sigma'}(\mathbf{V}\dot{\sigma}')+f\mathbf{K}\times\mathbf{V}+\nabla\Phi+\sigma'\alpha\nabla P^*=\mathbf{F}. \qquad \ldots (7)$$

The thermodynamic energy equation is

$$\frac{\partial}{\partial t}(C_pT)+\nabla\cdot(C_p T\mathbf{V})+\left(\frac{P}{P_{00}}\right)^{x}\frac{\partial}{\partial \sigma'}(C_p\dot{\sigma}'v)-\sigma'\alpha\left(\frac{\partial P^*}{\partial t}+\mathbf{V}\cdot\nabla P^*\right)=\dot{H} \qquad \ldots (8)$$

and the moisture continuity equations is

$$\frac{\partial}{\partial t}(q)+\nabla\cdot(q\mathbf{V})+\frac{\partial}{\partial \sigma'}(q\dot{\sigma}')=\dot{Q} \qquad \ldots (9)$$

and the mass continuity equation is

$$\nabla\cdot\mathbf{V}+\frac{\partial}{\partial \sigma'}(\sigma')=0 \qquad \ldots (10)$$

where

$$\nabla(P^*\mathbf{V}\mathbf{V})=(P^*\mathbf{V}\cdot\nabla)\mathbf{V}+\mathbf{V}\nabla+\mathbf{V}\nabla\cdot(P^*\mathbf{V})$$

$$\nabla(\mathbf{V}\mathbf{V})=(\mathbf{V}\cdot\nabla)\mathbf{V}+\mathbf{V}\nabla\cdot\mathbf{V}$$

with $P_{00}=1000$ hPa a reference pressure.

The forcing terms are:

$\mathbf{F}$ = horizontal frictional force vector per unit mass

$\dot{H}$ = diabatic heating rate per unit mass

$\dot{Q}$ = rate of moisture addition per unit mass

The primary dependent variables are:

$\boldsymbol{\Phi}$ = geopotential hight

α = specific volume
P = pressure

$$\dot{\sigma}=\frac{d\sigma}{dt},\ \dot{\sigma}_2=\frac{25}{P*}\nabla\cdot\mathbf{V}_5+\frac{1}{4P*}\nabla\cdot|P*(\mathbf{V}_3-\mathbf{V}_1)|,\ \text{Sigma vertical velocity}$$

$$\dot{\sigma}_4=\frac{50}{P*}\nabla\cdot\mathbf{V}_5\ ,\ \dot{\sigma}'_4=\nabla\cdot\mathbf{V}_5$$

$\mathbf{V}=(u, v)$, horizontal velocity vector

T = temperature

q = water vapour mixing ratio (kg/kg)

Physical Processes

The physical processes in the model include the conditions at the earth's surface, turbulent fluxes in the boundary layer, friction, diabatic heating and moisture sources, cloud types and radiations. The conditions at the earth's surface involve the orography, surface type, surface albedo, ground .temperature, ground hydrology, snow mass budget, snowmelt, monthly climatological sea-surface temperatures and sea ice. The turbulent fluxes are momentum, sensible heat and moisture fluxes. The diabatic heating consists of convective adjustment, large-scale condensation and evaporation cumulus convective. For the calculation of the radiation, the concentration of absorption gases in atmosphere, such as water vapour, carbon dioxide and ozone, are all constant, while the amount of high-, medium- and low-level clouds is variable.

The experimental schemes are as the following: (1) Control experiment (Exp. 1). The AGCM has been run for ten model years. We take the last fields from Nov. 1 to next Jul. 31 as the results of the control experiment. (2) Experiment on the anomaly snow cover over the Tibetan Plateau (Exp. 2). In this experiment, we run the AGCM from Nov. 1 to next summer but take the abnormal snow surface albedo over the Plateau between November and December. (3) Experiment without atmospheric diabatic heating over the Tibetan Plateau (Exp. 3). In this experiment, we run the AGCM from Apr. 1 to Jul. 31 but the atmospheric diabatic heating over the Plateau has been removed.

Analysis and Discussion of the Experimental Results

There are a low of experimental results obtained from the tests. In order to save space, the attention is focused on the differences of the air flow at upper and lower level, surface air temperature and precipitation in the Asian summer monsoon region so that we can examine the anomaly heating effects of the Tibetan Plateau on the Asian summer monsoon.

Results of the Control Experiment (Ex. 1)

In order to show the ability of the AGCM to reproduce the climate mean

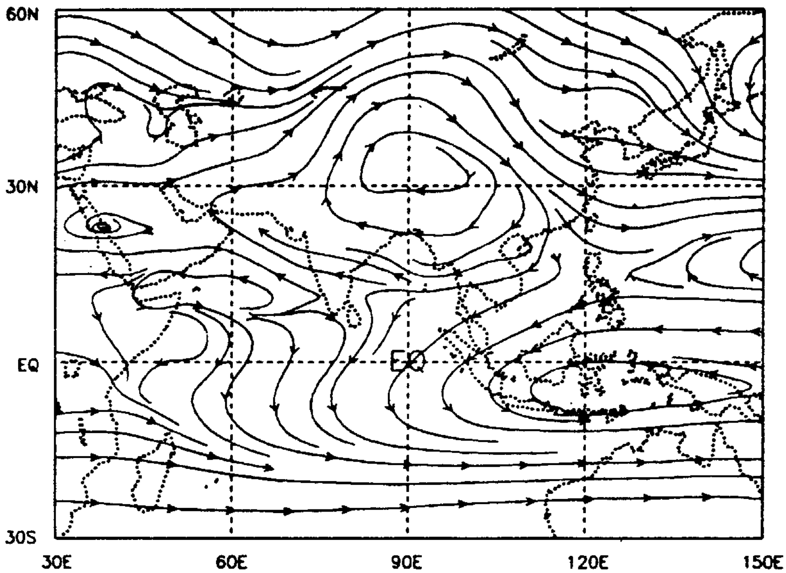

Fig 2(a) The simulated 250 hPa streamline in July

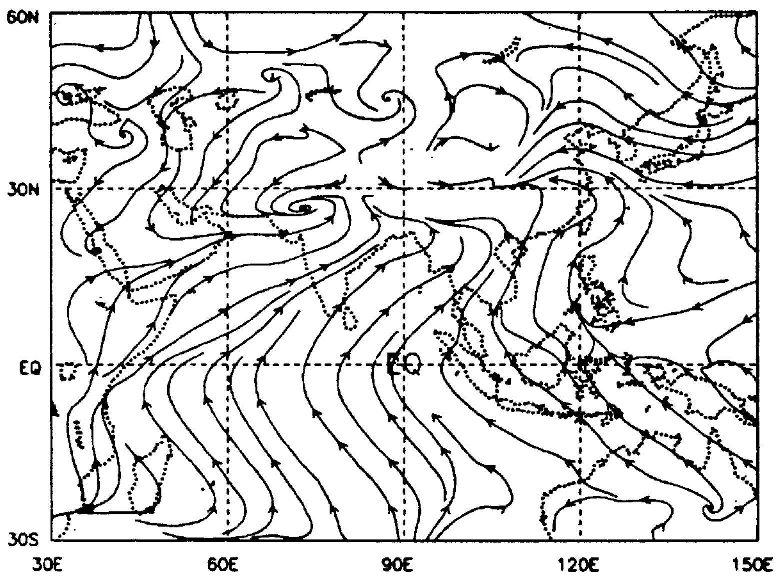

Fig 2(b) The simulated 975 hPa streamline in July

patterns of the Asian summer monsoon, we give some results from the control experiment.

The average July simulated flow patterns are shown in Figs 2*a* and 2*b*. At the upper level (250 hPa), the simulated southern Asian anticyclone is closer to that observed, its center is at 90°C, 32°N where the observed center is at

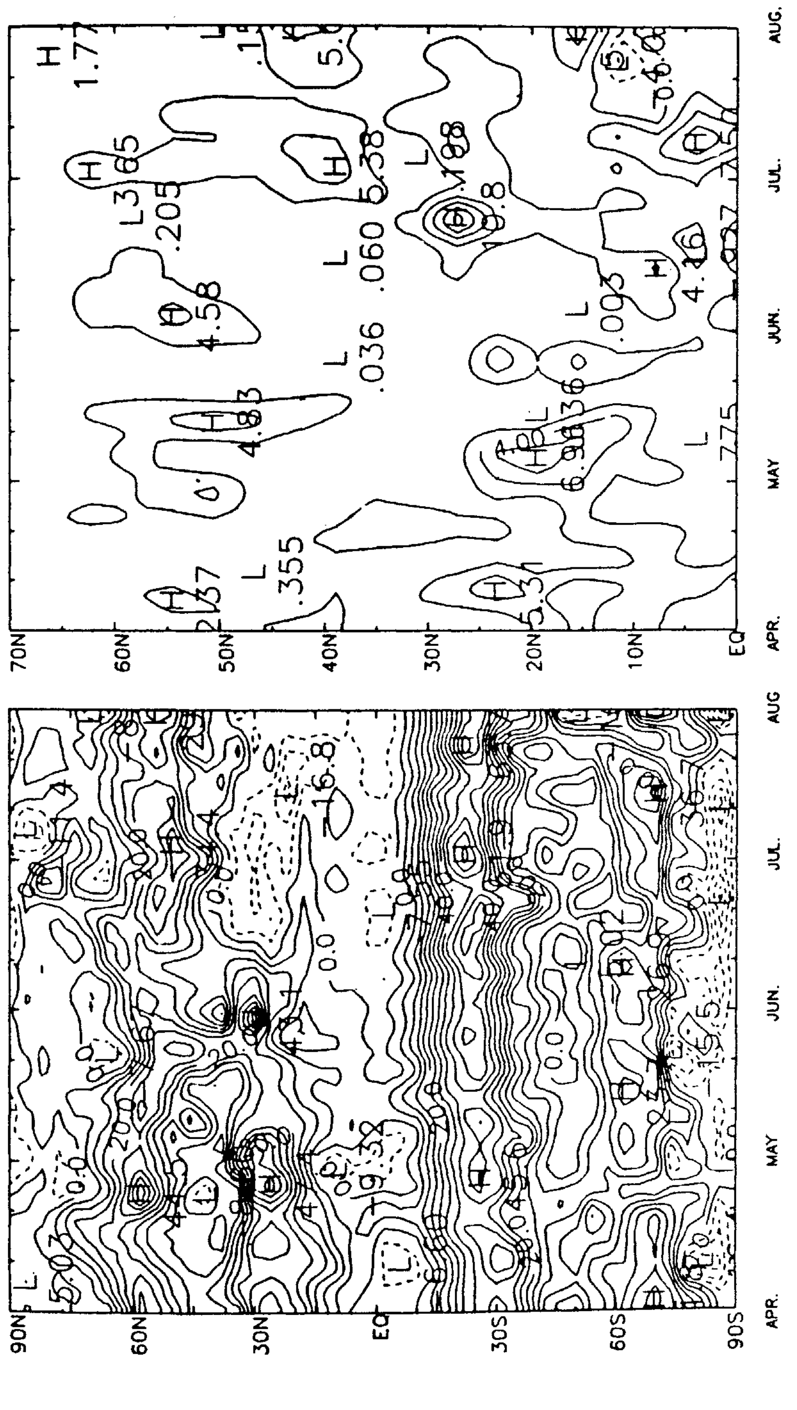

Fig 3 The simulated latitude-time cross-section of the (*a*) zonal wind (at 250 hPa) along 85-90°E and (*b*) precipitation along 110-120°E

88°E, 28°N. The simulated two easterly jets over the south side of Qinghai-Xizang Plateau, the northern one is linked with Asian upper anticyclone and the southern one is joined with the equatorial anticyclone and it is the most intense upper cross equatorial flow, are in better agreement with observation. At the lower level (975 hPa), the simulated low layer Somali jet, monsoon trough, the cross equatorial flows over the South China Sea and Bay of Bengal are in closer agreement with observation. The low pressure belt between Yangtze River and Huaihe River, the cold anticyclone over Australia are successfully simulated. The major features of simulated July surface air temperature, sea-level pressure and precipitation are in closer agreement with the observations.

It is an important symbol for a climate model to simulate the seasonal variation of the flow and precipitation successfully. The lititude-time cross section of the zonal wind (at 250 hPa) along 85-90°E is shown in Figs 3*a* and the latitude-time cross section of rainfal along 110-120°E is shown in Fig. 3*b*. From Figs 3*a*, we see that there are two westerly jets in northern and southern side of Tibetan plateau between April and May, one is the polar-front jet and the other is subtropical jet. Compared with its two sides, the zonal wind speed is weaker over the plateau. In early May, the easterly flow begins across the equator. In early June, the simulated subtropical westerly jet disappears suddenly and the easterly flow appears there immediately. In mid-summer there are two centres in the easterly flow. The result shows that this model can successfully simulate the seasonal variation of the flow at the upper level.

The simulated seasonal variation of precipitation from South China Sea to east of China is shown in Figs 3*b*. There are two rainfall belts, one is the ITCZ in South China Sea and the other is in the mainland. For the former, its centre is near the equator in May and moves northward to 15°N after middle July. This variation is in better agreement with the satellite observation. For the latter, it forms due to the interaction of the summer monsoon and cold air, and has the features of the subtropical rainfall belt. The observation shows that this rainfall belt is in South China during April to June and is called the rain belt in the former flood season (South China Maiyu), it jumps to the middle and lower Yangtze River region in middle June and is called Yangtze River Maiyu, and it again jumps to North China during July and August. From the Figs 3*b* we see that the simulated mainland rainfall belt is in South China (the centre near 22°N) during April to June and is in closer to the actual South China Maiyu, it jumps to near 30°N from middle June to July and is similar to the Yangtze River Maiyu, then it jumps again to 40°N in middle July and is in closer to North China rainfall belt. Therefore, the simulated seasonal variations are in better agreement with those observed in the rainfall belts in the mainland of China and the South China Sea.

From the analysis we can see that this model can simulate most features of the observation in the Asian summer monsoon region.

Results of Experiment on the Anomaly Snow Cover over the Tibetan Plateau

The anomaly snow cover over the Tibetan Plateau in winter will enhance the surface albedo in this area and make the surface heating effect decrease,

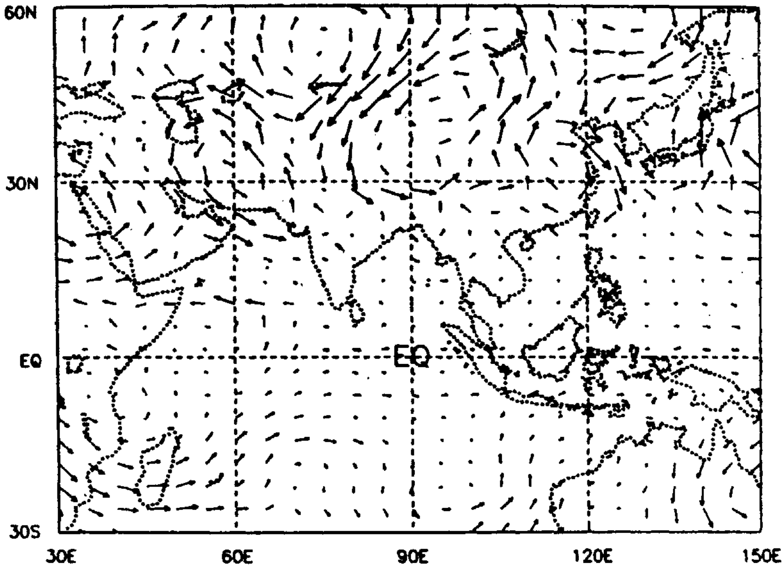

Fig 4(a) The difference of 250 hPa wind vectors in July

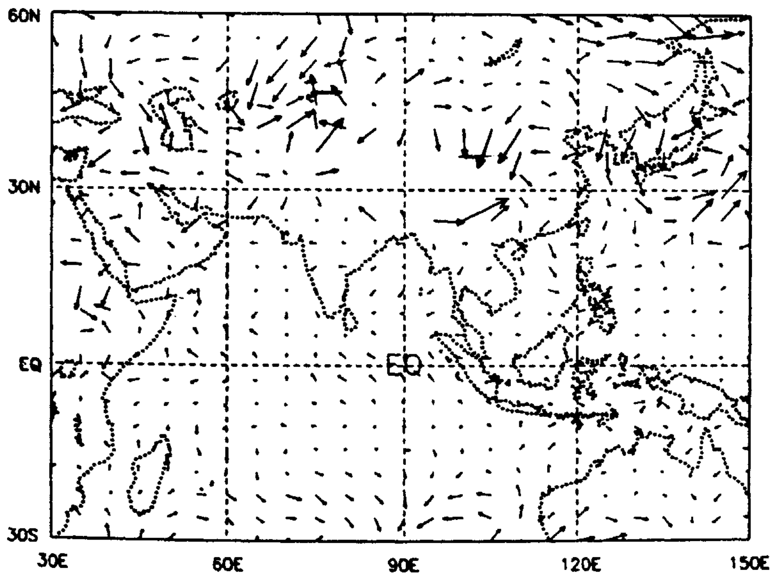

Fig 4(b) The difference of 975 hPa wind vectors in July

then cause the variation of circulation patterns. Now we want to test if this abnormal variation can influence the formation and evolution of the Asian summer monsoon circulation. Figs 4*a* and 4*b* are the differences of simulated monthly mean wind vectors at 250 hPa and 975 hPa level in July, respectively (Exp. 2 minus Exp. 1). We can see that at the upper level the anticyclone over the Plateau and west Asia has been weakened but it strengthened over East Japan. At the lower level, the SW monsoon flows in the Indian monsoon region and the South China Sea are weakened slightly, while a differential cyclonic circulation appears over the South-east Japan. From Figs 4*a* and 4*b* we can see that at the upper and lower level the differences of wind vectors are larger in middle and high latitudes in low latitudes and tropics.

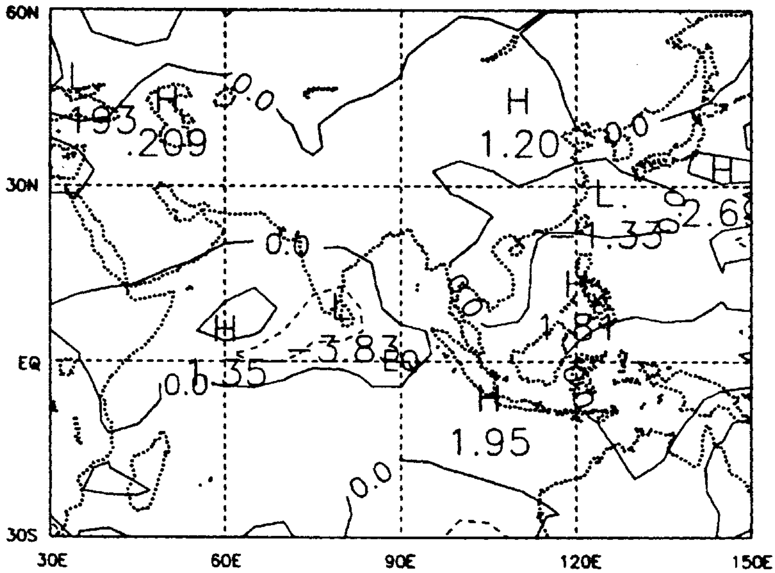

Fig 5 Difference of monthly mean precipitation in July (Exp. 2 minus Exp. 1).

The difference of the simulated monthly mean precipitation in July (Exp. 2 minus Exp. 1) is shown in Fig. 5. The observation show that if the snow cover over the Tibetan Plateau abnormally increased in winter, the precipitation in the middle and lower reaches of Yangze River will decrease and the mainland rainfall belt will shift southward in next summer[6]. From Fig 5, it is seen that the precipitation decreased in the middle and lower reaches of Yangze River and South India while it increased in North China, the South China Sea and East Japan. The simulated results are in closer agreement with those observed.

The result of the difference of the simulated monthly surface air temperature in July (Exp. 2 minus Exp. 1) shows that the surface air temperature is lower in most of China while it is higher in Indochina Peninsula and East Arabia Peninsula. The latitude-time cross sections of zonal wind and rainfall also show that the the polar jet over the north side of Plateau weakened and in east Asia the rainfall in Yangze River decreased but it increased in South China and South China Sea (Figs are omitted).

The above analysis shows that the response of the atmospheric general circulation to the abnormal snow cover over the Plateau is very sensitive, especially for the Indian summer monsoon system. Though this anomaly happens in winter, its effects can expend to the whole summer.

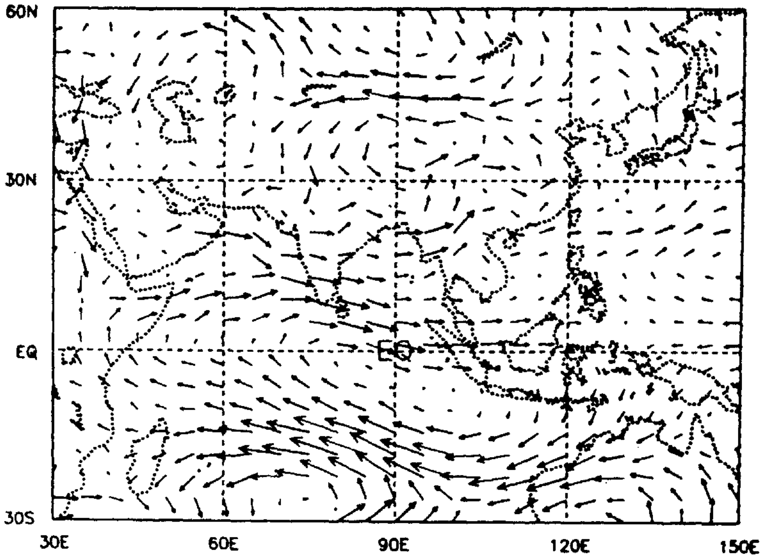

Fig 6(a) The difference of 250 hPa wind vectors in July

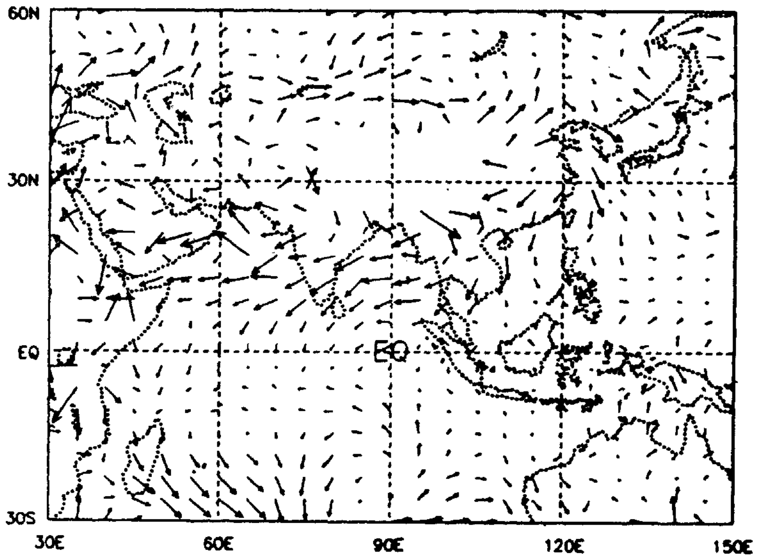

Fig 6(b) The difference of 975 hPa wind vectors in July

The atmospheric heat source anomaly over the Plateau in summer also has large effects on the circulation. The differences of the monthly mean wind vectors at 250 hPa and 975 hPa level in July are shown in Figs 6*a* and 6*b*, respectively (Exp. 3 minus Exp. 1). We can see that at the upper level, the South Asian anticyclone and the equatorial anticyclone are weakened obviously. Thus the upper easterly flows over the south Asia and cross-equatorial flows over the west of 90°E are also weakened significantly, while the anticyclone over the western Pacific is intensified clearly. At the lower level, there is a differential anticyclone from east Asia to west Asia. In the Indian summer monsoon area, the Somali jet and the SW monsoon flows over the Arabian Sea, the Indian subcontinent and Bay of Bengal are weakened obviously. The Indian monsoon trough disappeared and the cross-equatorial flows over the Indian Ocean are also weakened. In the east Asian monsoon area the SW monsoon flows over the East China and the north part of the South China Sea are weakened slightly. The south easterly flows over the western pacific are also weakened. From above analysis it is seen that the atmospheric diabatic heating has great influences on the Indian summer monsoon circulation, but for the east Asian summer monsoon system, its influence is not as obvious as that in the Indian monsoon system.

The difference of the simulated precipitation in July (Exp. 3 minus Exp. 1) is shown in Fig. 7. The figure shows that the precipitation increased in north India, south of Bay of Bengal and east of the South China Sea while it decreased in the Western Pacific, South India and East Arabia Peninsula. The results suggest that the atmospheric heating anomaly over the Plateau has important influence on the precipitation in low-latitude region of Asia.

The results of the latitude-time cross sections of the zonal wind at the upper level and the rainfall show that if there is no atmospheric diabatic heating over the Tibetan Plateau, the zonal wind speed over the Plateau and its neighbouring area will decrease obviously, its seasonal variation is not clear and at the upper level the easterly flows over the low-latitude of south Asia in summer disappear. In east Asia, the rainfall in the middle latitude of East China also decreased. In July the surface air temperature in most of Asia is lower, especially in East China and West India (Figs are omitted).

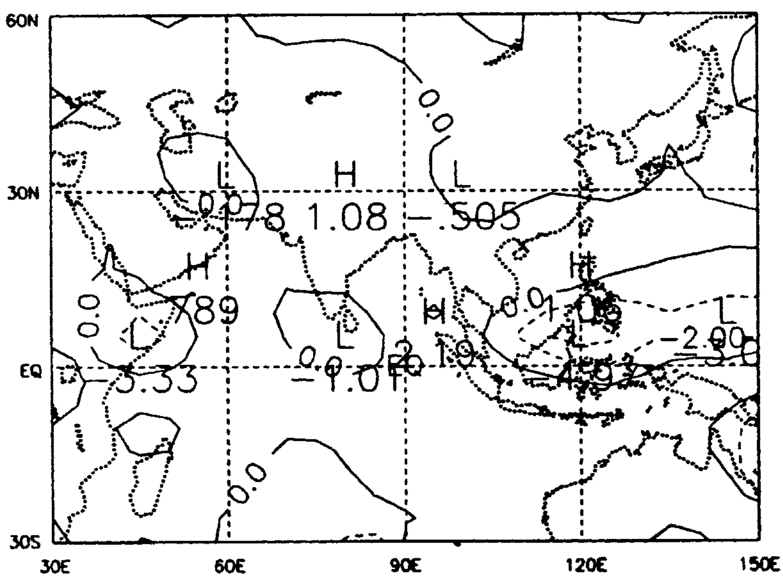

Fig 7 The difference of rainfall rate (mm/day) in July

The above discussion indicates that the atmospheric diabatic heating over the Tibetan Plateau is very important for the formation and evolution of Indian summer monsoon. Without this heating, the distribution of the heating fields over Asia will be changed and cause the variation of the Asian summer monsoon circulation: Indian summer monsoon disappear and the east Asian summer monsoon will be weakened.

Summary

Based on the above analysis and discussion, we can come to the following conclusions:

(1) The control experiment shows that the AGCM can reproduce most features of the observation in the Asian summer monsoon area.

(2) The effects of the anomaly snow cover over the Tibetan Plateau in winter can expend to the whole summer. A larger snow albedo than the normal on the Plateau in winter will weaken the summer monsoon in Asia. The precipitation in Yangze River will decrease and the rainfall belt in the mainland of China will shift southward which are in better agreement with those observed.

(3) The atmospheric diabatic heating over the Tibetan Plateau in summer is very important for the formation and evolution of the Asian summer monsoon circulation, especially for the Indian summer monsoon system. If this heating is removed, the Indian summer monsoon will disappear and the east Asian summer monsoon will be weakened.

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