NUMERICAL EXPERIMENTS ON THE FORMATION OF THE EASTERN PACIFIC SUMMER MONSOON

Gao Lei (高 雷),

Beijing Meteorological Bereau, Beijing 100081

Li Weiliang (李维亮)

Chinese Academy of Meteorological Sciences, Beijing 100081

and *Wang Bin* (王 斌)

Department of Meteorology, University of Hawaii at Manoa, USA

Received February 10, 1994; revised March 15, 1995

ABSTRACT

In this paper, we use a two-dimensional primary equation model which contains (1) heating of radiation, (2) heating of condensation, and (3) transfers of sensible and latent heat between air and the underlying surface. To investigate the causes for the formation of the eastern North Pacific summer monsoon, the data at 110°W are obtained and winds at underlying surface and at 200 hPa are modified under the conditions (1) removing topography and (2) changing meridional sea surface temperature (SST) gradient.

In the numerical modification, we find that by removing the topography, the center' s location of the eastern North Pacific summer monsoon does not change, but the intensity of the summer monsoon is weakened. Also the onset of the summer monsoon is delayed to the end of May. The tropical easterly jet is weakened obviously, even changes to westerly wind. On the other hand, we find that the SST gradient along 110°W influences the eastern North Pacific summer monsoon distinctly. If the SST gradient is decreased, the center of the southwest wind near 12°N does not exist any more, the intensity of the whole summer monsoon becomes very weak and the circulation pattern of the summer monsoon also changes a lot.

Finally, we indicate that both topography and meridional SST gradient play important roles in the occurrence of the eastern North Pacific summer monsoon. The meridional SST gradient is the most important factor that triggers the summer monsoon and the topography along 110°W influences the intensity and the onset time of the summer monsoon there mostly.

Key words: summer monsoon, sensitivity experiment, intensity, underlying surface

I. INTRODUCTION

In the recent years, research on monsoon has been developed greatly (Tao and Chen 1987). Among those works, the physical formation of monsoon is the main problem that the meteorologists explored. Many researchers (Halley 1986; Ramage 1971; Fein and Stevens. 1987) thought that the monsoon is totally driven by land-sea contrast and tried to explain the occurrence and the development of monsoon in Southeast Asia. But more and more meteorologists think many thermal and dynamical factors influence together on the occurrence of summer monsoon. For example, Zhang (1982) argued that the main characteristics of atmospheric circulation in Europe-Asia area result from the planet, sea-land and polar ice heat engines affecting each other. He pointed out that monsoon is the product of sea-land heat engine, but it is obviously only where the interaction of sea-land heat engine with the other two heat engines enhances; otherwise, it is very weak. Dickinson (1984) pointed out that the snow cover on the underlying surface has a good relation with the intensity of summer monsoon. Zhang et al. (1991) discussed the influence of vegetation on summer monsoon by using numerical experiment method.

Murakami et al. (1992) analyzed the formation of the summer monsoon over the Bay of Bengal and the eastern North Pacific respectively by using the data at 90°E and 110°W of the underlying wind, upper-level wind, surface pressure, SST and outgoing longwave radiation (OLR). They pointed out that the distinct sea-land contrast in East Asia is the main cause for the formation of summer monsoon at the Bay of Begnal. The influence of SST gradient over there is not important. However, this is not the case with the formation of the eastern North Pacific summer monsoon. Figure 1 shows the distribution of SST at 110°W. It is obvious that a northward SST gradient exists at the east part of Pacific and the time of this gradient forming is just in summer. After analyzing combined with the surface pressure data, they have found that it is the SST gradient that causes an obvious poleward surface pressure gradient, thus triggering and enhancing the northward cross-equatorial flow. They concluded that the physical factor on the formation of the eastern North Pacific summer monsoon is the meridional SST gradient at 110°W and the influence of topography is not important.

The eastern North Pacific summer monsoon is an important part of the global monsoon system. Murakami et al. (1992) explored the mechanism of the eastern North Pacific summer monsoon based on the objective data analysis method. In this paper, the numerical experiment method is used to investigate the mechanism of the eastern North Pacific summer monsoon.

II. THE EXPERIMENTAL PLAN

1. *Brief Description of the Model*

The two-dimensional global climate model developed by the Chinese Academy of Meteorological Sciences (CAMS) is used here. The atmospheric part of the model may refer to the paper by Li et al. (1989). The air velocity, temperature, moisture and surface pressure are computed based upon the equations of motion, the thermodynamic equation and the continuity equations for moisture and mass respectively. The heating of radiation, including the absorption, emission and scattering of the cloud, water vapour and carbon dioxide, is calculated by the schemes designed by Chen and Li (1985). The heating of condensation is divided into two kinds: one is caused by large-scale motion in the stable stratification and the other is by cumulus convection. The calculation for the former is according to research by Li et al. (1989) and latter according to the scheme of convective parameterization designed by Kuo (1974).

2. *The Experimental Design*

The lower boundary of the model is placed on the earth's surface along 110°W. And we let the area north of 25°N be land and south of 25°N be sea. The land surface temperature is

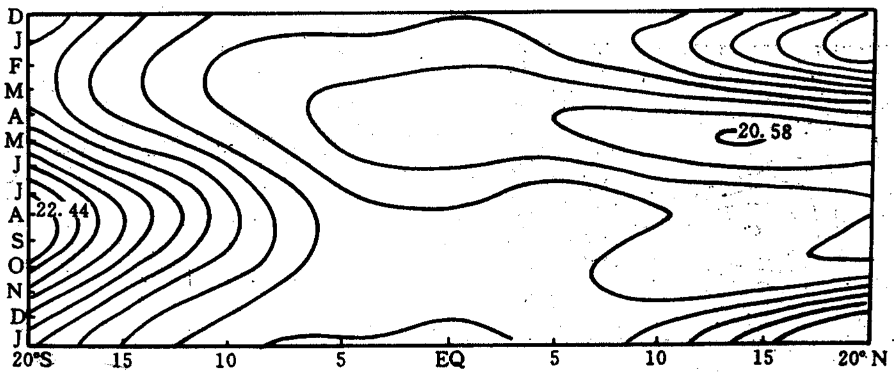

Fig. 1. The latitude-time distribution of SST along 110°W (interval 0.5° C) (from paper by Murakami et al. 1992).

calculated based on the heat balance equation at the underlying surface and the SST is monthly mean data.

First of all, a control experiment, in which the fixed land-sea boundary condition and the topography along 110°W are used, was carried out to provide a standard for sensitivity experiments. The topography data come from the data being used in the National Center for Atmospheric Research (NCAR) global model. The SST data are interpolated from the SST monthly mean data given in the paper by Murakami et al. (1992). And the SST data south of 25°S are still using the monthly mean SST data in the NCAR model. To investigate the influence of the topography and the SST on the monsoon respectively, two sensitivity experiments were performed. One was performed by removing the topography along 110°W without change of the other conditions, the other was by averaging the observational SST along 110° W only.

The numerical results became stable after the model had been integrated for three years both in the control and the two sensitivity experiments. The results in the third model year are used for discussions, with emphasis on the region 0—25°N in this paper.

III. THE EXPERIMENTAL RESULTS AND DISCUSSIONS

1. *The Control Experiment*

Figures 2, 3 and 4 show the observational zonal wind and meridional wind at the underlying surface and the observational zonal wind at 200 hPa respectively. It can be seen that southwest wind prevails in summer in 7—15° N with a maximum center while northeast wind is dominant in winter in the same region. At 200 hPa, northeast wind is dominant over 7—15°N in summer while southwest wind prevails in winter. Therefore from the observational wind field in upper and lower levels we can see that northeast wind prevails in winter along the south fringe of the North Pacific which is controlled by the sub-tropical high, the inter-tropical convergence zone (ITCZ) moves southward and southerly wind is dominant at the upper troposphere, thus Hadley circulation is enhanced; while in summer, the northeast wind changes into southwest wind at the underlying surface when the ITCZ (updraft leg of the Hadley circulation) reaches its northernmost latitude of about 14° N in July. The strong northeast wind prevails at the upper level and becomes the return flow of the southwest wind at the lower

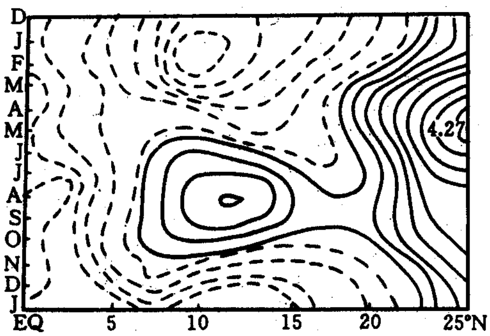

Fig. 2. The observation zonal wind at the underlying surface along 110°W (from paper by Murakami et al. 1992) (interval 1.0 m/s, dashed lines for easterly).

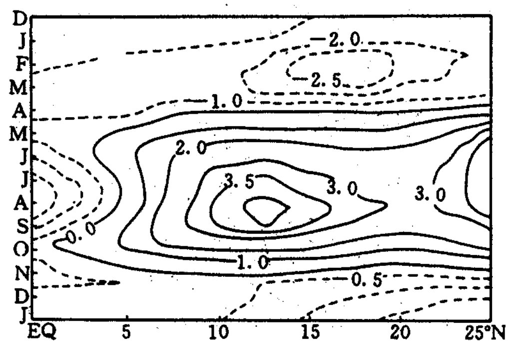

Fig. 5. As in Fig. 2 but for the control experiment (unit: m/s).

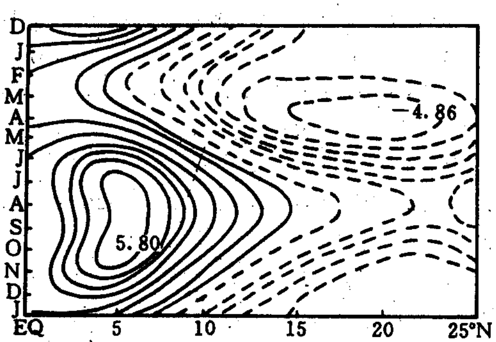

Fig. 3. As in Fig. 2 but for meridional wind (m/s).

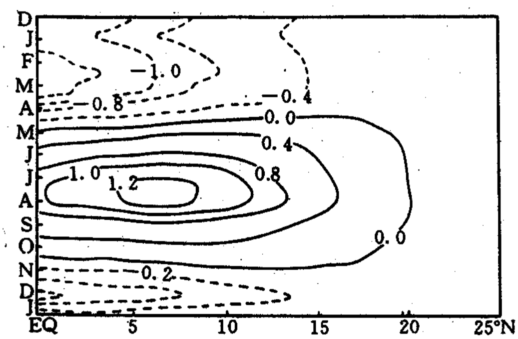

Fig. 6. As in Fig. 3 but for the control experiment (unit: m/s).

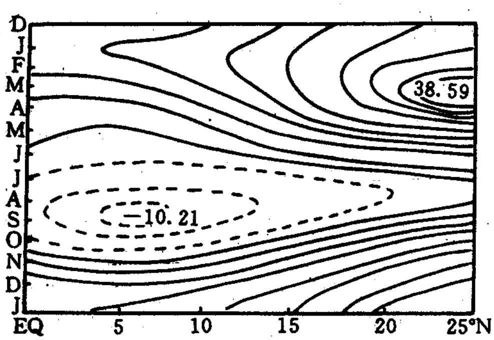

Fig. 4. As in Fig. 2 but for at 200 hPa (m/s).

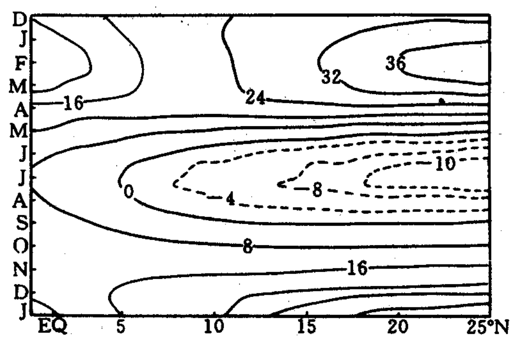

Fig. 7. As in Fig. 5 but for the control experiment (unit: m/s).

level. Thus, there actually exists a summer monsoon circulation system at the low latitude region over the eastern North Pacific.

Figures 5 to 7 show the zonal and meridional winds at the underlying surface and the zonal wind at 200 hPa simulated in the control experiment. We can see that the simulated main wind system at low latitudes in the Northern Hemisphere is almost the same as the observations. The simulated westerly center locates at about 12°N and appears in August, which is similar to the observations except for a little larger wind speed. In the control experiment, westerly wind prevails at high latitudes from spring to summer and the easterly wind controls the lower latitude area in the Northern Hemisphere. This simulated result is consistent with the observations while the intensity of air flow and the location of the maximum center are a little different from the observations. Northeast wind prevails at the region north of 20°N in winter in the control experiment, which is very different from the observations. Moreover, the simulated maximum center of south wind in summer locates at 5°N and appears in August, which is also similar to the observations except for a little less intensity. In winter, north wind prevails at lower latitude area in the Northern Hemisphere in the control experiment, which is similar to the observations, although the location of wind center is southward and the intensity is a little weak. At 200 hPa, the subtropical easterly jet and the westerly jet are also well simulated. The location of jets and their appearing time are basically consistent with the observations.

In general, the simulated winds are consistent with the observations, especially the zonal winds, which describe the occurrence and development of the eastern North Pacific summer monsoon system generally. Thus, the model used in the present paper can better simulate the variations of meteorological element fields and is useful for identifying physical mechanisms which control climate and its response to changes of various parameters.

2. *Sensitivity Experiment A—Without Topography*

Figure 8 is the zonal wind at the underlying surface in experiment A. After topography is removed, the underlying zonal wind system has not changed and there is still an obvious west wind center in August and September near 12°N, although the intensity is weakened somehow. The west wind appearing time delays from April, when the west wind appears in the control experiment, to the end of May. In 20—25°N, west wind prevails in summer in the control experiment, while it becomes weak east wind in experiment A. Moreover, the intensity of the whole summer monsoon in experiment A is weaker than that in control experiment.

Figure 9 is the zonal wind at 200 hPa in experiment A. After topography is removed,

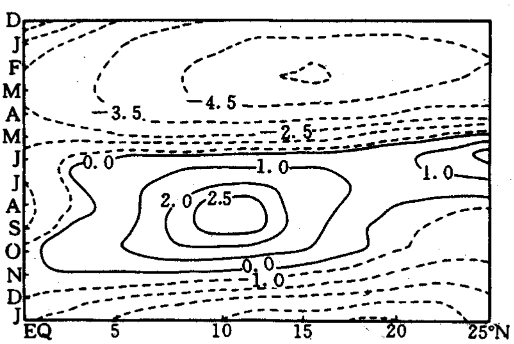

Fig. 8. As in Fig. 2 but for experiment A (m/s).

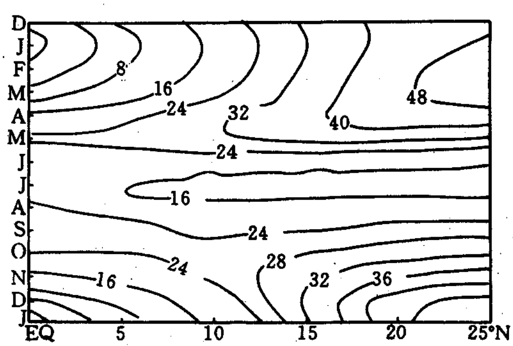

Fig. 9. As in Fig. 4 but for experiment A (unit: m/s).

west wind still prevails at upper level over the lower latitude region of the Northern Hemisphere in winter and its maximum center is still over 25° N and appears in February and March. The whole westerly belt is strengthened, its maximum value increased from 36 m/s to 48 m/s and its appearing time delays for half a month. The easterly belt, which appears in control experiment in July—August, is replaced by the weak west wind. Therefore removing topography enhances the west component of 200 hPa wind field greatly and weakens the upper easterly jet. It proves that the easterly jets both over eastern North Pacific and over East Asia are influenced by topography.

After topography is removed, the physical features of the underlying surface and the upper atmosphere will be changed. The following section will deal with physical process of variations.

Figure 10 shows the latitude-time distribution of difference of the underlying surface temperature between sensitivity experiment A and control experiment. As shown in this figure, the temperature in northern area (near 25°N) is lower than normal. Especially, there is a negative center in May whose maximum value is -4°C. In this period, the illuminating time of sunlight on the Northern Hemisphere begins to increase and the underlying surface temperature begins to go up. The negative temperature difference value here indicates that in and after spring, the underlying surface temperature is lower than normal and this lower underlying surface temperature lasts for a long time, so the underlying surface northward temperature gradient forms later than usual. In addition, a negative difference center appears after August, which maybe explain the variation of zonal wind at the underlying surface (in Fig. 8).

The variation of the underlying surface temperature has a close relationship with that of latent and sensible heat fluxes from the underlying surface to the atmosphere. Figure 11 is the same as Fig. 10 but for latent heat flux. Because the SST is not changed in this experiment, the variation of latent heat flux from sea surface to the atmosphere is not distinct. It is obvious that the latent heat released by the moisture in the ascending air decreases greatly because of the low temperature on the land. The negative center is near 20°N which appears in May—August when summer monsoon prevails. The sensible heat flux difference (figure not shown) also indicates increasement of sensible heat flux from the underlying surface to the atmosphere

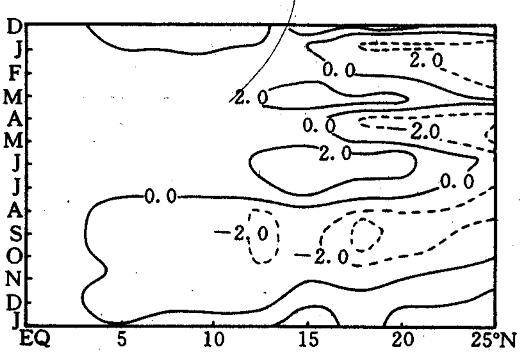

Fig. 10. The latitude-time distribution of difference of the underlying surface temperature between the sensitivity experiment A and the control experiment (°C).

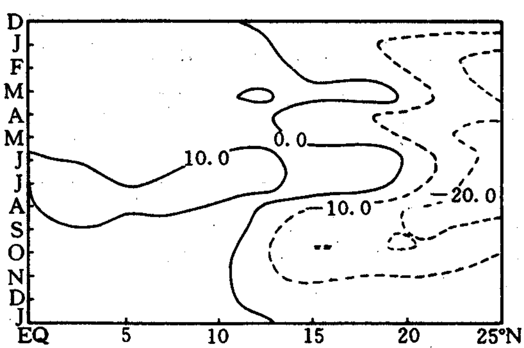

Fig. 11. As in Fig. 10 but for latent heat flux (W/m^2).

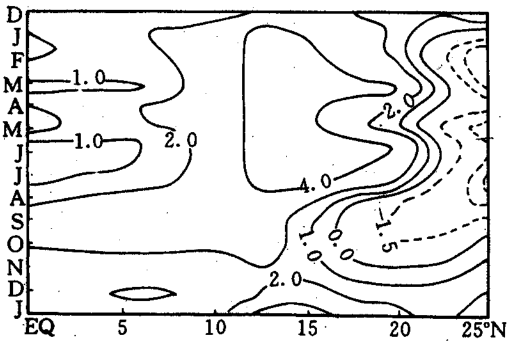

Fig. 12. As in Fig. 10 but for the temperature difference at 900 hPa (℃).

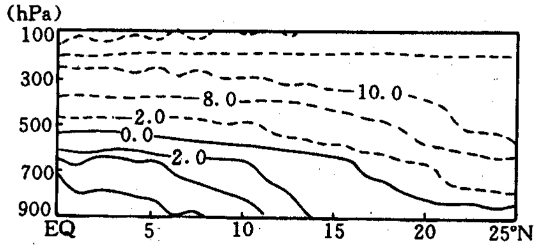

Fig. 13. The latitude-altitude distribution of mean temperature difference between the sensivity experiment A and the control experiment in May (℃).

because of the low temperature on land.

The variation of the sensible and latent heat from the underlying surface to the atmosphere and the variation of the long wave radiation have great influences on the temperature structure and circulation pattern of the atmosphere. Figure 12 is the same as Fig. 10 but for the temperature difference at 900 hPa. As shown in this figure, the atmospheric temperature at 900 hPa over the land decreases after removing topography and its maximum center appears in May. On the contrary, the atmospheric temperature at 900 hPa over the sea near the equator tends to increase. Thus the meridional difference of the atmospheric temperature at lower levels decreases relatively. This is not benefit to the formation and development of the northward temperature gradient which triggers the summer monsoon. Figure 13 is the latitude-altitude distribution of mean temperature difference between sensitivity experiment A and control experiment in May. During the spring, the seasonal variation of physical features and wind fields is the most distinct. As noted in this figure, the atmospheric temperature in the middle troposphere (about 300 hPa) over the sea increases while the temperature at the upper troposphere decreases obviously. Then, the northward temperature gradient in the middle troposphere over the southside of topography increases relatively. This causes the tropical easterly jet to weaken, even to disappear.

3. *Sensitivity Experiment B*

To make a further investigation to the mechanism of the summer monsoon at low latitude region in the eastern North Pacific, we keep the topography along 110°W as in the control experiment and meanwhile average the observational SST along 110°W yearly to reduce the meridional SST gradient.

Figure 14 is the zonal wind at the underlying surface in experiment B. After averaging the SST yearly, the most evident changes are that the maximum west wind center near 12° N disappears, the seasonal variation of the underlying surface wind is not distinct, and the summer monsoon becomes very weak. The intensity of the west wind in this region becomes weak obviously. By the end of summer, west wind prevails in 20—25°N, which is consistent with that in the control experiment.

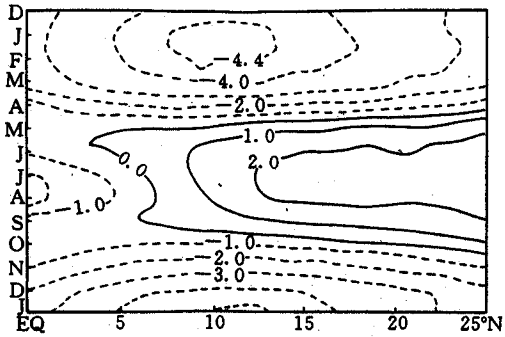

Fig. 14. As in Fig. 8 but for experiment B (m/s).

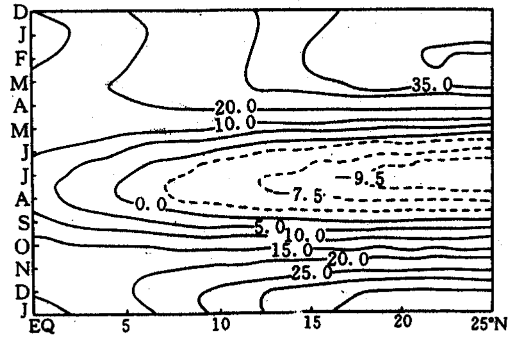

Fig. 15. As in Fig. 9 but for experiment B (m/s).

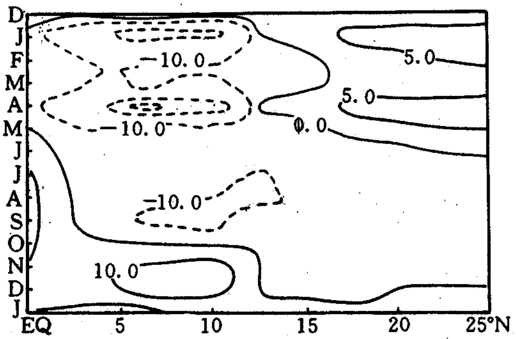

Fig. 16. As in Fig. 11 but for experiment B (W/m²).

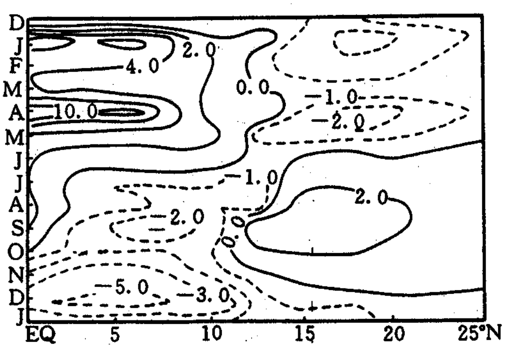

Fig. 17. As in Fig. 16 but for sensible heat flux (W/m²).

Figure 15 is the zonal wind at 200 hPa. The intensity of the easterly jet in summer becomes weak. In winter, the intensity of the 200 hPa westerly jet increases a little. In general, the influence of meridional SST gradient on the 200 hPa easterly jet is less important than that of topography.

Since the SST along 110°W is yearly averaged and has not monthly variation, the SST in the eastern equatorial Pacific cold water region becomes warmer than normal from spring to summer while the SST in the region north of the equator decreases, and the northward SST gradient becomes very weak. Then the meridional underlying surface temperature gradient, which drives the summer monsoon, has not existed. This distinct variation of the meridional SST gradient results in variations of sensible and latent heat fluxes from the surface to the atmosphere.

Figures 16 and 17 are latitude-time distributions of difference of the latent and sensible heat fluxes between sensitivity experiment B and control experiment respectively. As shown in Fig. 16, the latent heat from the ascending air over sea in 10—15°N to the surrounding atmosphere decreases greatly while the latent heat received by the atmosphere over the equator increases. In addition, there is also a negative center of the latent heat near 10°N in April. The results above show that since April, the meridional temperature gradient becomes very weak although the SST begins to go up because of the growing sunlight, the convergence of the ascending air is very weak and the latent heat from the warm wet air over the sea decreases

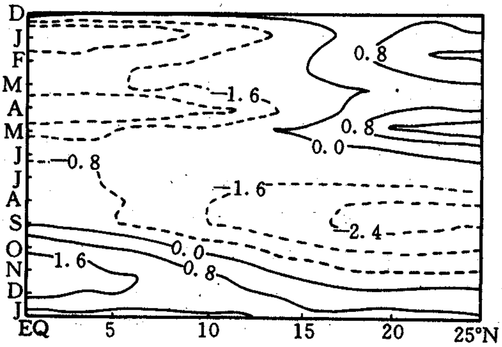

Fig. 18. As in Fig. 16 but for atmospheric temperature at 900 hPa (°C).

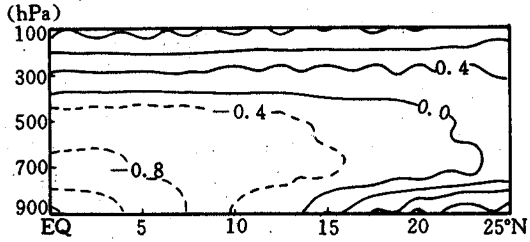

Fig. 19. As in Fig. 13 but for experiment B (°C).

greatly. On the contrary, the latent heat from the sea to the atmosphere near the equator from spring to summer increases, the descending motion of atmosphere is weakened, even disappeared. This is also indicated in Fig. 17. This condition will last until the end of summer and will influence the atmospheric temperature structure and circulation pattern.

Figure 18 is the same as Fig. 16 but for atmospheric temperature at 900 hPa. As shown in this figure, the atmospheric temperature at the lower troposphere decreases from spring to summer with an extreme value in August over 10—15°N while the temperature over the equator increases. Then the meridional atmospheric temperature gradient becomes weak, which influences greatly on the formation and development of the summer monsoon. Figure 19 is the same as Fig. 13 but for experiment B. The atmospheric temperature in the mid troposphere decreases both over equator and over 10—15°N. But the magnitude is small, showing the little variation of the meridional temperature gradient in the mid troposphere and little influence of the SST gradient on the tropical easterly jet.

IV. SUMMARY

The results of numerical experiments above prove the viewpoint that the eastern North Pacific summer monsoon is primarily induced by SST contrast over the ocean. This viewpoint adds and develops the theory that summer monsoon is mainly driven by sea-land contrast. But in the mean time we also point out that the influence of topography can not be ignored. As we discussed above, topography plays a very important role in the onset and termination time and the intensity of the summer monsoon. It also has a close relationship with the tropical easterly jet. To some extent, topography is one of the physical factors in the formation of the eastern North Pacific summer monsoon. It is incorrect to ignore the influence of topography on the summer monsoon.

Thus, both topography and meridional SST gradient play very important roles in the formation of the eastern North Pacific summer monsoon. The meridional SST gradient is the decisive factor on the formation of the summer monsoon. After the SST is averaged yearly, the monsoon becomes very weak and the center of the surface southwest wind does not exist. The topography along 110°W plays an important role in the intensity, onset and break time of the monsoon. It also has influence on occurrence of the tropical easterly jet. Without topography, the monsoon occurs later and breaks early, and the intensity of the tropical easterly jet decreases obviously, even becomes westerly wind.

REFERENCES

Chen Longxun and Li Weiliang (1985), The atmospheric heat budget in summer over Asia monsoon area, *Adv. Atmos. Sci.*, **2**: 487 —496.

Dickinson, R. E. (1984), Modelling evapotranspiration for three-dimensional global climate model, Climate processes and climate sensitivity, Geophysical Monograph 29, Maurice Ewing Volume 5, J. E. Hansen and T. Takahashi, Eds., American Geophysical Union, Washington, D. C., pp. 58—72.

Fein, J. and Stevens, P. (1987), *Monsoons*, John Willey and Sons, New York, pp. 632.

Halley, E. (1686), A historical account of the trade winds and monsoons observable in the seas between and near the tropics with an attempt to assign the physical cause of the said wind, *Phill. Trans. Roy. Soc. London*, **16**: 153—168.

Kuo, H. L. (1974), Further studies of the parameterization of the influence of cumulus convection on large-scale flow, *J. Atmos. Sci.*, **31**: 1232—1240.

Li Weiliang, Chen Longxun, Zhou Xiuji and Gong Wei (1989), The numerical simulation of two-dimensional dynamical climate model with mountain forcing, *Acta Met. Sin.*, **3**: 312—327.

Ramage, C. S. (1971), *Monsoon Meteorology*, Academic Press, New York, pp. 427.

Tao, S. Y. and Chen, L. X. (1987), *The East Asian Summer Monsoon in Reviews in Monsoon Meteorology*, Oxford University Press, pp. 60—92.

Murakami, T., Wang, B. and Lyons, S. W. (1992), Contrasts between Summer Monsoons over the Bay of Bengal and the Eastern North Pacific, *J. Meteor. Soc. Jap.*, **70**: 43—62

Zhang Jiacheng (1982), *The Seasonal Variation of the Atmosphere Circulation and the Scientific Concepts of Monsoon*, Documents of the National Tropical Summer Monsoon Meeting, China Meteor. Press, pp. 225.

Zhang Zhenqiu, Li Weiliang and Chen Longxun (1991), Numerical Experiments on the Effect of Qinghai-Xizang, *Acta Met. Sin.*, **5**: 442—454