ABRUPT CHANGE OF FLOW PATTERN IN BAROCLINIC ATMOSPHERE FORCED BY JOINT EFFECTS OF DIABATIC HEATING AND OROGRAPHY

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ABSTRACT

Based on the catastrophic theory, the possible causes of abrupt change in the atmospheric circulation over the Northern Hemisphere during June and October have been explored by Li and Luo (1983) and Miao and Ding (1985). However these studies are confined to the barotropic atmosphere without consideration of orography. The purpose of this paper is to further study the physical mechanism of the abrupt change of flow pattern within the baroclinic atmosphere in the presence of orography. Results show that the abrupt change of flow pattern can be stimulated by the gradual variation of a diabatically heating parameter, which is similar to the observed fact about the rapid shift of position of the subtropical high center in the upper troposphere along the zonal direction during seasonal transition from the summer half year to the winter one.

I. INTRODUCTION

Abrupt change in the flow pattern of large-scale quasi-geostrophic current is one of the basic forms of motion in the atmospheric circulation system. It is a very spectacular problem to probe into the physical mechanism of this basic motion form as well as those of other basic motion forms, such as the periodic and chaotic ones. With the low-order spectral model, some results regarding the physical mechanism of abrupt change in flow pattern have been obtained (Li and Luo, 1983; Liu and Tao, 1983; Miao and Ding, 1985). However, these studies are restricted to the barotropic atmosphere without taking orography into account. Whether the results for the barotropic atmosphere can still hold good for the baroclinic one is a problem to be solved. Here we introduce the baroclinicity and orography into the model used by Li and Luo (1983), in which the dependent variables of corresponding nonlinear system increase from four to twelve. The main purpose of this paper is to study the physical causes of abrupt change of flow pattern in the 12th order nonlinear system, forced and dissipative, which governs the dynamics of the baroclinic quasi-geostrophic current forced by the joint effects of diabatic heating and orography.

II. MODEL

The subscripts 1, 2, and 3 are used to denote the levels of 250, 500, and 750 hPa respectively. Thus, the geostrophic vorticity equations can be written at levels 1 and 3, and the thermal equation at level 2 as follows:

$$\frac{\partial}{\partial t}\nabla^2\psi_1 = -J(\psi_1, \nabla^2\psi_1 + \beta^* y) + f_0\frac{\omega_2}{\Delta P} - K_0'\nabla^2(\psi_1 - \psi_3), \tag{1}$$

$$\frac{\partial}{\partial t}\nabla^2\psi_3=-J(\psi_3,\nabla^2\psi_3+\beta^*y)-f_0\frac{\omega_2}{\Delta p}-K_6'\nabla^2(\psi_3-\psi_1)$$

$$-K_d\nabla^2\psi_3-\frac{f_0}{H}J(\psi_3,h), \tag{2}$$

$$\frac{\partial}{\partial t}(\psi_1-\psi_3)=-J(\psi_2,\psi_1-\psi_3)+\frac{f_0}{\Delta p}\lambda^2\omega_2-\lambda^2h_d'[(\psi_1-\psi_3)^*-(\psi_1-\psi_3)], \tag{3}$$

where ψ is the geostrophic streamfunction, ω the individual pressure change, h the lower boundary elevation, and the other symbols are of conventional meanings. The model atmosphere is confined to a periodic beta-plane channel with zonal walls at $y=0$ and $y=\pi D$, $D=1.83\times10^8$ cm; at the central line of the beta-plane, $\phi_0=40°N$.

Let

$$\psi_1=\psi+\theta,\quad \psi_3=\psi-\theta,\quad \psi_2=\psi,$$

and

$$(\psi,\theta,\omega)=\sum_i(\psi_i,\theta_i,\omega_i)F_i,$$
$$\theta^*=\theta_A^*F_A+\theta_C^*F_C,\qquad h=h_KF_K. \tag{4}$$
$$(i=A,K,L,C,M,N)$$

The expressions of F_i can be referred to the paper by Charney and Straus (1980).

Substituting (4) into the dimensionless forms of Eqs. (1)—(3), we have the spectral system:

$$\dot\psi_A=-K_2(\psi_A-\theta_A)+\frac{1}{2}h_{01}(\psi_L-\theta_L), \tag{5}$$

$$\dot\psi_K=-\beta\alpha(\psi_A\psi_L+\theta_A\theta_L)-\beta\alpha''(\psi_C\psi_N+\theta_C\theta_N)+\beta_1\psi_L-K_2(\psi_K-\theta_K), \tag{6}$$

$$\dot\psi_L=\beta\alpha(\psi_A\psi_K-\theta_A\theta_K)+\beta\alpha''(\psi_C\psi_M+\theta_C\theta_M)-\beta_1\psi_K-K_2(\psi_L-\theta_L)-\frac{h_{n1}}{2}(\psi_A-\theta_A), \tag{7}$$

$$\dot\psi_C=\epsilon\alpha''(\psi_K\psi_N+\theta_K\theta_N)-\epsilon\alpha''(\psi_L\psi_M+\theta_L\theta_M)-K_2(\psi_C-\theta_C)-\frac{h_{02}}{2}(\psi_N-\theta_N), \tag{8}$$

$$\dot\psi_M=-\delta\alpha''(\psi_C\psi_L+\theta_C\theta_L)-\delta'\alpha'(\psi_A\psi_N+\theta_A\theta_N)+\beta_2\psi_N-K_2(\psi_M-\theta_M), \tag{9}$$

$$\dot\psi_N=\delta\alpha''(\psi_C\psi_K+\theta_C\theta_K)+\delta'\alpha'(\psi_A\psi_M+\theta_A\theta_M)-\beta_2\psi_M-K_2(\psi_N-\theta_N)-\frac{h_{n2}}{2}(\psi_C-\theta_C), \tag{10}$$

$$\dot\theta_A=A_{11}(\psi_K\theta_L-\psi_L\theta_K)+A_{12}(\psi_M\theta_N-\psi_N\theta_M)+A_{13}\psi_A-A_{14}\theta_A+A_{15}\theta_A^*$$
$$-h_{A1}(\psi_L-\theta_L), \tag{11}$$

$$\dot\theta_K=K_{11}\alpha\psi_L\theta_A+K_{11}\alpha''\psi_N\theta_C-K_{12}\alpha\psi_A\theta_L-K_{12}\alpha''\psi_C\theta_N+K_{13}\psi_K-K_{14}\theta_K+K_{15}\theta_L, \tag{12}$$

$$\dot\theta_L=-K_{11}\alpha\psi_K\theta_A-K_{11}\alpha''\psi_M\theta_C+K_{12}\alpha\psi_A\theta_K+K_{12}\alpha''\psi_C\theta_M+K_{13}\psi_L-K_{14}\theta_L$$
$$-K_{15}\theta_K+h_{L1}(\psi_A-\theta_A), \tag{13}$$

$$\dot\theta_C=C_{11}\alpha''(\psi_M\theta_L-\psi_N\theta_K)+C_{12}\alpha''(\psi_K\theta_N-\psi_L\theta_M)+C_{13}\psi_C-C_{14}\theta_C$$
$$+C_{15}\theta_C^*-h_{C1}(\psi_N-\theta_N), \tag{14}$$

$$\dot\theta_M=M_{11}\alpha''\psi_L\theta_C-M_{12}\alpha''\psi_C\theta_L+M_{13}\alpha'\psi_N\theta_A-M_{14}\alpha'\psi_A\theta_N+M_{15}\psi_M$$
$$-M_{16}\theta_M+M_{17}\theta_N, \tag{15}$$

$$\dot\theta_N=M_{11}\alpha''\psi_K\theta_C+M_{12}\alpha''\psi_C\theta_K-M_{13}\alpha'\psi_M\theta_A+M_{14}\alpha'\psi_A\theta_M+M_{15}\psi_N$$
$$-M_{16}\theta_N-M_{17}\theta_M+h_{N1}(\psi_C-\theta_C), \tag{16}$$

where $K_2=\frac{K_1}{2}$, $\alpha=\frac{40\sqrt{2}\,n}{15\pi}$, $\alpha'=\frac{32\sqrt{2}\,n}{15\pi}$, $\alpha''=\frac{64\sqrt{2}\,n}{15\pi}$,

$$\beta=\frac{n^2}{n^2-1},\ \beta_1=\frac{\beta^* n}{n^2-1},\beta_2=\frac{\beta^* n}{n^2-4},\ \delta=\frac{n^2-3}{n^2+4},\delta'=\frac{n^2+3}{n^2+4},$$

$$\epsilon=\frac{3}{4},h_{01}=\alpha h_K,\ h_{02}=\frac{\alpha''}{4}h_k,\ h_{n1}=\frac{h_{01}}{n^2+1},\ h_{n2}=\frac{\alpha'' h_K}{n^2+4},$$

$$h_{A1}=\frac{h_{01}\sigma_0}{2(\sigma_0+2)},h_{L1}=\frac{h_{n1}\sigma_0}{2L_{11}},L_{11}=2(1-\beta)+\sigma_0,\ B_{11}=\frac{1}{\sigma_0+2},$$

$$h_{C1}=\frac{h_{02}\sigma_0}{2D_{11}},D_{11}=2(1-\epsilon)-\sigma_0,\ h_{N1}=\frac{h_{n2}\sigma_0}{2N_{11}},N_{11}=2(1-\delta')+\sigma_0,$$

$$A_{11}=\frac{2\alpha}{B_{11}},A_{12}=\frac{2\alpha'}{B_{11}},A_{13}=\frac{K_2\sigma_0}{B_{11}},A_{14}=\frac{K_2\sigma_0+2K_v\sigma_0+2h''}{B_{11}},$$

$$A_{15}=\frac{2h''}{B_{11}},\ K_{11}=\frac{2(1-\beta)-\beta\sigma_0}{L_{11}},\ K_{12}=\frac{2(1-\beta)+\beta\sigma_0}{L_{11}},\ K_{13}=\frac{K_2\sigma_0}{L_{11}},$$

$$K_{14}=\frac{K_2\sigma_0+2K_v\sigma_0+2(1-\beta)h''}{L_{11}},\quad K_{15}=\frac{\beta_1\sigma_0}{L_{11}},C_{11}=\frac{2(1-\epsilon)-\epsilon\sigma_0}{D_{11}},$$

$$C_{12}=\frac{2(1-\epsilon)+\epsilon\sigma_0}{D_{11}},C_{13}=\frac{K_2\theta_0}{D_{11}},C_{14}=\frac{K_2\sigma_0+2K_v\sigma_0+2(1-\epsilon)h''}{D_{11}},$$

$$C_{15}=\frac{2(1-\epsilon)h''}{D_{11}},\ D_{11}=2(1-\epsilon)+\sigma_0,\ h''=g_1\sigma_0,\ M_{11}=\frac{2(1-\delta')-\delta\sigma_0}{N_{11}},$$

$$M_{12}=\frac{2(1-\delta')+\delta\sigma_0}{N_{11}},\quad M_{13}=\frac{2(1-\delta')-\delta'\sigma_0}{N_{11}},\quad M_{14}=\frac{2(1-\delta')+\delta'\sigma_0}{N_{11}},$$

$$M_{15}=\frac{K_2\sigma_0}{N_{11}},M_{16}=\frac{K_2\sigma_0+2K_v\sigma_0+2(1-\delta')h''}{N_{11}},\ M_{17}=\frac{\sigma_0\beta_2}{N_{11}},N_{11}=2(1-\delta')+\sigma_0,$$

n is the zonal wavenumber within the beta-plane. The following parametric values are prescribed: $K_v=0.0114$, $K_2=0.0057$, $h''=0.0114$, and $n=2$, which are the same as those in the papar by Charney and Straus (1980). Let $\sigma_0=0.109$, meaning that $2\sigma_S/\Delta P=25.7$ K/500 hPa^{-1}, which equals 86% of the typical value (Yoden, 1983). In addition, h_K is taken to be 0.09.

Put $\varphi=(\psi_A,\psi_K,\psi_L,\psi_C,\psi_M,\psi_N,\theta_A,\theta_K,\theta_L,\theta_C,\theta_M,\theta_N)$. It is easily seen from Eqs. (5)—(16) that $\varphi=\bar{\varphi}_0=(\psi_{A1},0,0,\psi_{C1},0,0,\bar{\theta}_{A1},0,0,\bar{\theta}_{C1},0,0)$ is the equilibrium solution of the spectral system.

The expressions of ψ_i and $\bar{\theta}_i$ (i=A,C) are:

$$\psi_{A1}=\bar{\theta}_{A1}=\frac{h''}{(K_v\sigma_0+h'')}\theta_A^*, \tag{17}$$

$$\psi_{C1}=\bar{\theta}_{C1}=\frac{(1-\epsilon)h''}{K_v\sigma_0+(1-\epsilon)h''}\theta_C^*. \tag{18}$$

According to the perturbational equation system corresponding to (5)—(16), the characteristic equation has the form:

$$\lambda I-A=0, \tag{19}$$

where I is a unit matrix, A is a 12×12 coefficient matrix.

Eq. (19) will be used to determine whether the equilibrium states defined by (17)—(18) are stable.

III. INSTABILITY OF THE FLOW PATTERN OF HOMOGENEOUS DISTRIBUTION OF GEOPOTENTIAL HEIGHT ALONG SUBTROPICAL HIGH AND TWO KINDS OF STABLE SUBTROPICAL FLOW PATTERNS

By use of formulas (17)—(18) the equilibria $\bar{\varphi}_0$ of (5)—(16) are computed. Eighty-eight pairs of the values in the range of both $0.02\leqslant\theta_A^*\leqslant0.09$ and $-0.055\leqslant\theta_C^*\leqslant-0.005$ are impartially selected for the specified values of θ_A^* and θ_C^*. These equilibria $\bar{\varphi}_0$ represent the flow patterns of homogeneous distribution of geopotential height along subtropical high-pressure belt. By substituting the various equilibria $\bar{\varphi}_0$ into Eq. (19), the various characteristic roots of (19) may be obtained. The computational results on the characteristic roots show that the equilibria corresponding to the eighty-eight pairs of the values of θ_A^* and θ_C^* are all instable, which illustrates that the flow patterns of homogeneous distribution of geopotential height along subtropical high are all instable. These results are similar to the observed facts in the real atmosphere. It will be shown that this instability can stimulate two kinds of stable subtropical flow patterns.

Experiment 1

Put $\varphi=\varphi_{01}=(0.045, 0, 0, -0.021, 0, 0, 0.045, 0, 0, -0.021, 0, 0)$ as an initial value. The integration for Eqs. (5)—(16) is carried out for the period of 375 model days. The time step is taken to be 3 h, and the computational scheme from Asselin (1972) (hereinafter, the time step and computational scheme are the same as those in Exp. 1). The values of θ_A^* and θ_C^* are determined by (17)—(18) and φ_{01}, and remain constant in the process of the integration. Having integrated Eqs. (5)—(16) to the 48th model day or later, we find that the closed subtropical high center at the upper troposphere is always located over the south of model plateau, forming a quasi-stationary motion form. For the sake of convenience, we define the range of $0\leqslant x<\frac{\pi}{4}$, $\frac{3\pi}{4}\leqslant x<\frac{5\pi}{4}$, and $\frac{7\pi}{4}\leqslant x<2\pi$ as model land, and the range of $\frac{\pi}{4}\leqslant x<\frac{3\pi}{4}$. $\frac{4}{5\pi}\leqslant x<\frac{4}{7\pi}$ as model sea.

Experiment 2

Put $\varphi=\varphi_{02}=(0.0676, 0, 0, -0.020, 0, 0, 0.0676, 0, 0, -0.020, 0, 0)$ as an initial value; the values of θ_A^* and θ_C^* are also determined by (17)—(18) and φ_{02}, and remain constant in the process of the integration. The integration of Eqs. (5)—(16) is also carried out for the period of 375 model days. Starting from the 42th model day, it remains the state in which the closed subtropical high center over the south of the plateau oscillates around the orographic ridge in the range of $-\frac{\pi}{8}\leqslant x\leqslant\frac{\pi}{8}$ along the zonal direction. The oscillation has a non-attenuating amplitude with time and a period about two weeks, forming another form of motion with stable quasi-periodic variation.

In order to graphically show the evolution character of flow pattern, we let

$$F_k=\sum_i|(\psi_i+\theta_i)_k-(\psi_i+\theta_i)_{400}|,$$

$$(i=A,K,L,C,M,N,k\geqslant400)$$

where k denotes the ordinal number of the time step. If $F_k=0$, the situation of the geopotential height field at 250 hPa for time step k is much more similar to that for time step 400. When F_k achieves its maximum, the difference between the two situations is quite obvious.

Experiments 3—5.

We take $\theta_A^*=0.05$, and $\theta_C^*=-0.0285$, and define $\varphi=\varphi_{03}=$(0.03729, −0.02224, 0.00489, −0.01270, 0.02076, −0.00298, 0.02881, −0.01136, 0.00513, −0.00222, 0.00989, −0.002467) as an initial value (Exp. 3). The integration of Eqs. (5)—(16) shows that the variation of the system is characterized by quasi-periodic oscillation (Fig. 1b). In the meantime, in any phases of the curve in Fig. 1b, the corresponding subtropical high centers at 250 hPa are all located over the south of model plateaus rather than over model seas. By taking $\theta_A^*=0.05$, $\theta_C^*=-0.0290$, and the initial value $\varphi=\varphi_{03}$, the integration of the system is performed (Exp. 4). From the result of Exp. 4, we can see that the system enters into one stable equilibrium state (Fig. 1a). The 250 hPa streamfunctional field averaged over 340—370 model days of Exp. 4 (Fig. 2a) manifests that the subtropical high centers are located in the longitudes of model plateaus, and we can regard this flow pattern as "plateau" pattern. By taking $\theta_A^*=0.05$, and $\theta_C^*=0.0290$, and the initial value $\varphi=\varphi_{04}$ (Exp. 5), the computational result shows that the system enters into another stable equilibrium (Fig. 1c). The 250 hPa streamfunction field averaged over 720—750 model days of Exp. 5 (Fig. 2b) exhibits that the subtropical highs are located over model seas. Similarly, we regard this flow pattern as "sea" pattern.

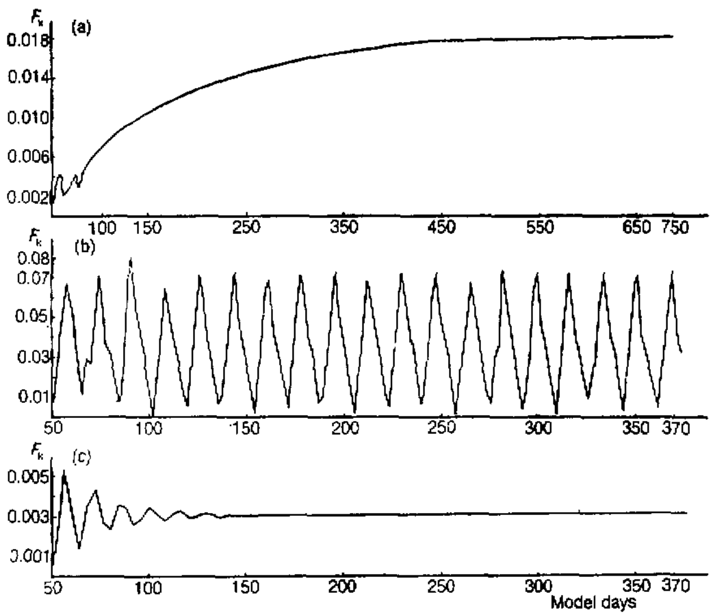

Fig. 1. Variations of F_k with time. $\theta_A^*=0.05$. (a) $\theta_C^*=-0.0290$, the inital value $\varphi=\varphi_{03}$; (b) $\theta_C^*=-0.0285$, the initial value $\varphi=\varphi_{03}$; and (c)$\theta_C^*=-0.0290$, the initial value $\varphi=\varphi_{04}$.

In addition, we have also performed tens of numerical experiments on the set of (5)—(16) under the circumstances of both various thermal forcing parameters and various initial conditions. The results are similar to those in Exps. (1—5).

To sum up, the flow patterns of homogeneous distribution of geopotential height along subtropical high belt are all certainly instable. After the flow patterns lose their stabilities, two kinds of motional forms may be created depending on both various initial values and various forcing parameters. One of them is the transient quasi-periodic variation, which is characterized by low-frequency oscillation of subtropical high around the longitude of the ridge line of model plateau along the zonal direction. The other belongs to the quasi-stationary waves. Under the same forcing effects, on condition that the initial values are taken to be different, the quasi-stationary position of subtropical high may be over either model plateau or model sea. Thus, the multiplicity of the equilibrium of the system can be exhibited clearly.

Inspecting the "plateau" pattern and "sea" pattern (Fig. 2) in a meticulous way, we find that the positions of the subtropical high corresponding to the two flow patterns are near those in the real atmosphere. The observed facts show that the subtropical high center is often located over the North Pacific in the winter half year, but over the longitudes of the Tibetan Plateau in the summer half year. Either in the time range of the winter half year or in that of the summer one, the position of the high center changes gradually and slightly. However, in the period of transition between these two half years, the position of the high center changes abruptly and obviously (Fig. 3). This is another example about abrupt change of the atmospheric circulation.

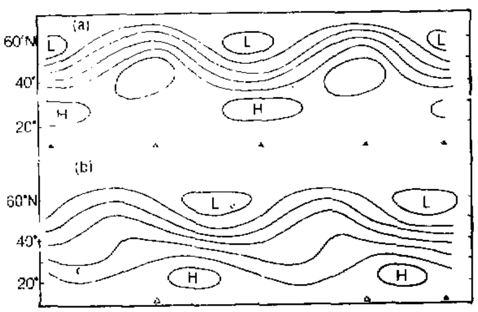

Fig. 2. The averaged streamfunction fields at 250 hPa. $\theta_A^*=0.05$, $\theta_C^*=-0.0290$. (a) "plateau" pattern; and (b) "sea" pattern. ▲ and Δ represent the longitudinal positions of the ridge lines of model plateau and the central lines of model sea, respectively. The intervals of isopleths equal 4×10^{-3}.

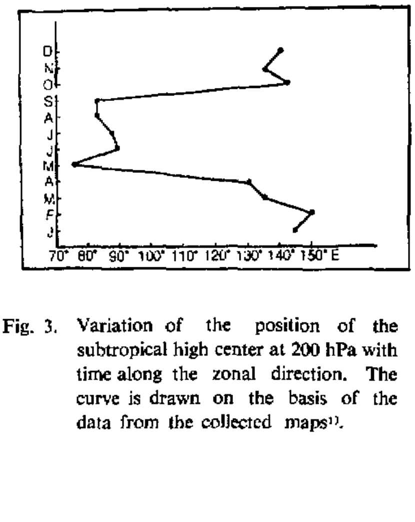

Fig. 3. Variation of the position of the subtropical high center at 200 hPa with time along the zonal direction. The curve is drawn on the basis of the data from the collected maps[1].

IV. GRADUAL OR ABRUPT CHANGE OF THE POSITION OF SUBTROPICAL HIGH CENTER AT THE UPPER TROPOSPHERE

Take $\theta_A^*=0.05$, and $\theta_C^*=-0.05$, -0.04, -0.035, -0.0325, -0.030, -0.0295, -0.0290, -0.0289, -0.0288, -0.0287, ..., -0.0281, which correspond to point a, b, c, d, e, f, g,..., o in Fig. 4. Starting from the initial field of "plateau" pattern, the integration of Eqs. (5)—

1) Institute of Central Meteorological Bureau (1972), *Aerological Atlas of the Northern Hemisphere*.

(16) is performed until the 375th model day. The results of sixteen integrations show that the subtropical high centers at 250 hPa are always located over the south of model plateau in the process of the integrations. From point a to point o, θ_c^* changes gradually from -0.05 to -0.0281. This roughly denotes a transition period from midsummer to autumn. In the transition period the intensity of the high center decreases gradually, but its position remains quasi-stationary.

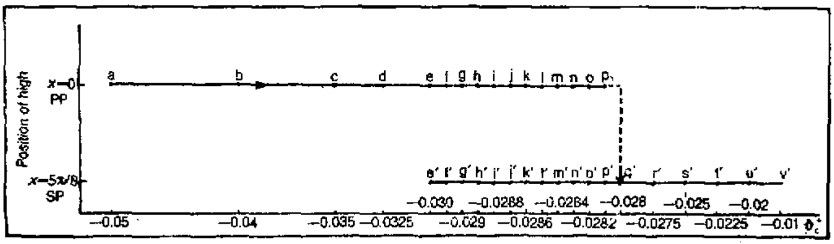

Fig.4. Schematic diagram of the abrupt change of the flow pattern in the transition period from midsummer to winter.

Let $\theta_c^* = -0.0280$, which corresponds to point p in Fig. 4, and the initial value $\varphi = \varphi_{03}$ ("plateau" pattern). The integration of the system (5)—(16) is carried out for a period of 375 model days. The result shows the system enters into such a state where the subtropical high center is persistently located over model sea rather than over the south of model plateau.

When a point shifts from o to p, the variation of $|\theta_c^*|$ is only 1/219 of that occurring when the point shifts from a to o, but a fundamental variation takes place in the system state. Therefore the form of motion of the abrupt change in flow patterns is clearly exhibited

Here, the unidirectional time property of the state change is also clear. Letting $\theta_c^* = -0.0281$, and starting from the initial field of the "sea" pattern φ_{04} rather than of the "plateau" pattern φ_{03}, the integration of Eqs. (5)—(16) is also performed for 375 model days. The flow pattern existing in the process of the integration is still the situation which is much more similar to the "sea" one. Similarly, taking the initial value $\varphi = \varphi_{04}$, and letting $\theta_c^* = -0.0282$, -0.0283, $-0.0284, \ldots, -0.0300$, which correspond to the points o′, n′, m′, l′, ..., e′ in Fig.4 or letting $\theta_c^* = -0.0275$, -0.0250, -0.0200, -0.0100, which correspond to the points r′, s′, t′, u′, v′, in Fig. 4, various integrations of Eqs. (5)—(16) are also carried out for 375 model days. All these computational results show that the subtropical high centers are still persistently located over the model sea.

In the range of $-0.030 \leqslant \theta_c^* \leqslant -0.028$, there are two stable equilibria of the flow patterns. Taking $\theta_c^* = -0.0300$, -0.0290, -0.0289, and starting from the "plateau" pattern φ_{03}, the integration of Eqs. (5)—(16) shows that the system enters into the quasi-periodic motion form in which the low-frequency oscillations are made of the subtropical high center around the longitude of the ridge line of model plateau along the zonal direction with the quasi-periods of 16, 18, and 20 model days, respectively. Under the same thermal forcing and starting from the "sea" pattern φ_{04}, the system enters into the quasi-stationary motion form in which the subtropical high center is persistently located over the model sea. Taking $\theta_c^* = -0.0288$, $-0.0287, \ldots, -0.0281$, and the initial values $\varphi = \varphi_{03}$ or φ_{04}, the system gets into the quasi-stationary "plateau" flow pattern or the quasi-stationary "sea" flow pattern.

Therefore, the integrations of the system (5)—(16) show (1) that in the certain parametric combinations, the system possess two stable equilibria certainly, (2) that the equilibria corresponding to the flow patterns of homogeneous distribution of geopotential height along the subtropical high pressure belt are all instable, and (3) that the time concept of evolution of the system both before and after the abrupt change possesses the unidirectional character. These are just consistent with the three conditions about the abrupt change of the atmospheric circulation (Li and Luo, 1983).

V. SUMMARY

Previous studies regarding the physical mechanism of the abrupt change of the atmospheric circulation by means of catastrophic theory are confined to the barotropic atmosphere without consideration of orography. The number of dependent variables of the corresponding forced, dissipative, and nonlinear systems is not more than six. The main purpose of this paper is to study the possible causes of the abrupt change of the flow pattern within a baroclinic atmosphere in the presence of orography. The model used in this paper is a twelfth order system of non-linear ordinary differential equations, which is deduced from the two-layer quasi-geostrophic model equations by the way of low-order spectral model. By using the model equations, tens of numerical experiments whose integrational time is longer than 370 model days have been carried out. The results show that along with the gradual change of the thermal forcing parameter the abrupt change of the subtropical flow pattern at the upper troposphere can clearly be exhibited. This is similar to the observed fact about rapid shift of the position of the subtropical high at the upper troposphere along the zonal direction during the period of seasonal transition from the summer half year to the winter one in the real atmosphere.

The states of the system both before and after the abrupt change within the barotropic atmosphere are all the equilibrium ones (Li and Luo, 1983). After the baroclinic process enters into the model, the states before the abrupt change of the flow pattern from the summer half year to the winter one are of the quasi-periodic form besides the equilibrium states. The periodic motion can be characterized by low-frequency oscillations of the subtropical high center around the longitude of the ridge line of model plateau. The time scales corresponding to the quasi-periodic form of motion about 2—3 weeks. This form of motion is different from either the stationary waves or the baroclinic waves traveling in westerly belt. It seems that this is a new form of transient waves, and may be related to the observed fact about the east-west oscillation of the subtropical high at the upper troposphere during the summer half year.

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