The Role of Topography and Diabatic Heating in the Formation of Dipole Blocking in the Atmosphere

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ABSTRACT

In this paper, the nonlinear stationary waves forced by topography and diabatic heating are investigated. It is pointed out that (1) the nonlinear interaction of different stationary waves forced only by topography might form dipole blocking in the atmosphere, this might explain the dipole blocking appeared in the Pacific and Atlantic regions; (2) the dipole blocking could not be caused by the nonlinear interaction of the different stationary waves forced by the diabatic heating alone; (3) the nonlinear interaction of the diffferent stationary waves forced by both topography and diabatic heating could initiate dipole blocking in the atmosphere. In winter, the dipole blocking mainly occurs in the west regions of the Pacific and the Atlantic, and the heat source over the western part of the two oceans is advantageous to the formation of dipole blocking in the west of two oceans. However, in summer, the dipole blocking could be formed in the east part of the two oceans, and the heat source over the eastern part of two continents is favourable for the formation of dipole blocking in the east regions of two oceans.

I. INTRODUCTION

In the atmospheric science, the blocking dynamics has been an important subject of the medium–range weather forecast. In the recent years, a series of blocking theories have been proposed by different authors. For example, the multiple equilibria theory by Charney and Devore (1979), and this theory can explain the global blocking. However, it can not reveal the localized structures of dipole blocking in the atmosphere. Later, Mcwilliams (1980) proposed the equivalent modon theory for local blocking, and the equivalent modon exhibits the vortex pair structure which is similar to dipole blocking in the atmosphere, but it can not describe the stationary property of dipole blocking. Recently, Luo and Ji(1988) proposed an algebraic Rossby solitrary wave theory of dipole blocking, which can reveal the stationary property of dipole blocking, but the evolution process of dipole blocking can not be described. And then, the authors (1987) suggested that the envelope Rossby solitary wave could describe the evolution of dipole blocking. In fact, some observational facts demonstrate that dipole blocking usually appears in the Pacific and Atlantic. Clearly, the dipole blocking relates to the forcings of topography and diabatic heating. Tung and Lindzen (1979a) showed that atmospheric blocking could be explained by the linear resonance of planetary scale waves with respect to surface forcing such as continental elevation and land–sea differential heating. Nevertheless, this theory can not reveal the nonlinear process of blocking formation and explain the dipole blocking in the Pacific and the Atlantic. In this paper, our main purpose is to study these problems.

II. THE BASIC EQUATION

For inviscid flow, the quasi-geostrophic potential vorticity equation with topography and diabatic heating in a β channel can be written as

$$\frac{\partial}{\partial t}\nabla^2\psi + J(\psi, \nabla^2\psi + \frac{f_0}{H}h) + \beta\frac{\partial\psi}{\partial x} = -bQ. \quad (1)$$

Eq.(1) is the equivalent barotropic potential vorticity equation, where ψ represents the geostrophic stream function, H is the equivalent depth of the atmosphere, h denotes the topographic elevation, f_0 is the Coriolis parameter at a central latitude, $\nabla^2 = \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2}$ denotes the horizontal Laplace operator, $J(a,b) = \frac{\partial a}{\partial x}\frac{\partial b}{\partial y} - \frac{\partial a}{\partial y}\frac{\partial b}{\partial x}$, $\beta = \frac{\partial f}{\partial y}$ is the gradient of Coriolis parameter at a central latitude, and bQ denotes the diabatic heating.

Introduce the non-dimensional variables in the following forms

$$(x,y) = L(\bar{x},\bar{y}), \quad \psi = UL\bar{\psi}, \quad Q = Q_0\bar{Q}, \quad h = h_0^*\bar{h}, \quad (2)$$

where $\bar{x}$, $\bar{y}$, $\bar{\psi}$, $\bar{Q}$ and $\bar{h}$ are the non-dimensional variables, U, L and h_0^* denote the characteristic wind speed, the characteristic length and the characteristic height of topography respectively, and Q_0 represents the characteristic value of planetary scale diabatic heating.

Substituting Eq.(2) into Eq.(1) and dropping "−", we may obtain the non-dimensional potential vorticity equation in the from

$$\frac{\partial}{\partial t}\nabla^2\psi + J(\psi, \nabla^2\psi + \frac{F}{R_0}h) + \beta_0\frac{\partial\psi}{\partial x} = -BQ, \quad (3)$$

where $F = h_0^* / H$, $R_0 = U/(f_0 L)$, $\beta_0 = \beta L^2 / U$, $B = bQ_0 L^2 / U$.

Let $\psi = -y + \psi'(x,y)$, Eq.(3) can be rewritten as follows:

$$\frac{\partial}{\partial x}\nabla^2\psi' + \beta_0\frac{\partial\psi'}{\partial x} + (F/R_0)\frac{\partial h}{\partial x} = -BQ - J(\psi', \nabla^2\psi' + (F/R_0)h) \quad (4)$$

where ψ' is the stationary disturbance stream function.

III. THE ROLE OF TOPOGRAPHIC FORCING

When the thermal forcing is not considered in Eq. (4), that is, $Q = 0$, Eq.(4) becomes

$$\frac{\partial}{\partial x}\nabla^2\psi' + \beta_0\frac{\partial\psi'}{\partial x} + (F/R_0)\frac{\partial h}{\partial x} = -J(\psi', \nabla^2\psi' + (F/R_0)h). \quad (5)$$

In Eq. (4), we may take $H \sim 10$ km, $U \sim 10$ m/s, $f_0 \sim 10^{-4}$ 1/s, $L \sim 2200$ km and $h_0^* \sim 300$ m. In this case $F/R_0 = 0.66$, and we can consider F/R_0 as a small parameter and let $\varepsilon = F/R_0$.

Here, we may expand ψ' asymptotically in terms of the small parameter ε of the form

$$\psi' = \varepsilon\psi_1 + \varepsilon^2\psi_2 + \ldots \quad (6)$$

Substituting Eq.(6) into Eq.(5), we have

$$O(\varepsilon^1): \frac{\partial}{\partial x}\nabla^2\psi_1 + \beta_0\frac{\partial\psi_1}{\partial x} = -\frac{\partial h}{\partial x}. \quad (7)$$

$$O(\varepsilon^2): \frac{\partial}{\partial x}\nabla^2\psi_2 + \beta_0\frac{\partial\psi_2}{\partial x} = -J(\psi_1, \nabla^2\psi_1 + h). \quad (8)$$

In this study, we may assume that the topography consists of M mode harmonic waves, and the topography distribution is assumed to be

$$h = \frac{1}{2}\sum_{l=1}^{M} h_l \exp[ilkx]\sin ny + cc, \tag{9}$$

where cc denotes the complex conjugate of the preceding term, h_l is the amplitude of topographic wave, and $h_l \gg h_{l-1}$ ($l = 1 \cdots M$). In our computation, we may let $k=2$, $n=1$, and two mode orographic waves are assumed to exist in the atmosphere, i.e. $M=2$, thus, Eq.(9) can be reduced to

$$h = (h_1 \cos kx + h_2 \cos 2kx)\sin ny. \tag{10}$$

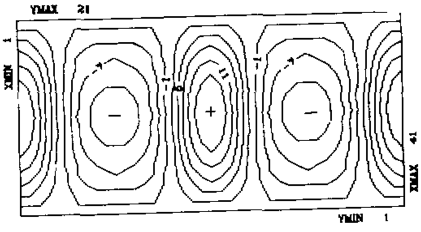

Fig. 1. Topography distribution, $h_1 = 1.5$, $h_2 = 0.3$, interval: 3, unit:0.1.

Let $h_1 = 1.5$, $h_2 = 0.3$, then, the topography distribution is shown in Fig. 1, where "+" represents the continents and "−" represents the oceans. Therefore, Fig.1 is equivalent to the idealized topography distribution of the two oceans and two continents which roughly reflects the topography distribution of Northern Hemisphere, it mainly consists of topographic waves 2 and 4. While if the nonlinear term of Eq.(5) is neglected, and Eq.(10) is introduced into Eq.(5), we may obtain the linear disturbance stream function solution for the stationary wave forced by topography, i.e.

$$\psi' = -\varepsilon\left[\frac{h_1}{\beta_0 - (k^2 + n^2)}\cos kx + \frac{h_2}{\beta_0 - (4k^2 + n^2)}\cos 2kx\right]\sin ny. \tag{11}$$

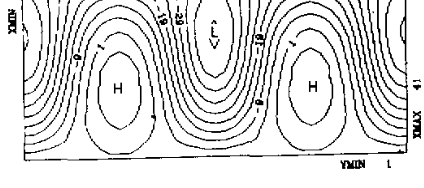

Fig. 2. The stream function field of linear response to the forcing of orography alone. interval: 5, unit: 0.1.

The flow field of stationary wave induced by topography is plotted in Fig.2. It is found that a blocking high which resembles blocking situation observed in two oceans appears in the valley region of topography.

Charney and Devore (1979) studied the multiple equilibrium states forced by topography and obtained one stable equilibrium state which is similar to Fig. 2. Therefore, from above discussion, we may see that the linear topographically forced stationary wave can basically characterize the blocking situations occurred in two oceans, but, it can not explain some particular sutructures of blocking situations over two oceans. A typical example of these particular structures is dipole blocking formed in the Pacific and the Atlantic. Frederiksen (1982) made a numerical experiment about the formation of blocking in the atmosphere, he found that the dipole blocking mainly occurred in two oceans. Some observational studies also showed this point, for example, the observational study of O'conner (1963). He also found that a vortex pair blocking with a high pressure north to a low appeared in the northeast Atlantic ocean during January 1963. Obviously, explanation to these phenomena is highly desired. In order to explain the dipole blocking in the atmosphere, Mcwilliams(1980) proposed an equivalent modon theory to interpret the blocking with vortex pair structure in the atmosphere. Although the equivalent modon can describe the local structure of a dipole blocking to a certain degree, the equivalent modon can not explain the seasonal and geographical variation of dipole blocking in the atmosphere, particularly for its geographical preference over the ocean. Observation shows that the dipole blocking usually occurs in different regions of the two oceans in different seasons. Clearly, this relates to topography and land–sea contrast. Next, we shall discuss the role of topography in the formation of dipole blocking in the atmosphere.

Inserting Eq. (10) into Eqs.(7) and (8), the nonlinear solution of the stationary wave forced by topography can be obtained in the following form.

$$\psi' = -\varepsilon\left[\frac{h_1}{\beta_0-(k^2+n^2)}\cos kx+\frac{h_2}{\beta_0-(4k^2+n^2)}\cos 2kx\right]\sin ny$$
$$+\varepsilon^2\frac{3}{4}nh_1h_2\sin 2ny\cos kx\left\{-\frac{3k^2}{[\beta_0-(k^2+n^2)]^2[\beta_0-(4k^2+n^2)]}\right.$$
$$\left.+\frac{2}{[\beta_0-(k^2+n^2)]^2}-\frac{2}{[\beta_0-(k^2+n^2)][\beta_0-(4k^2+n^2)]}\right\}. \qquad (12)$$

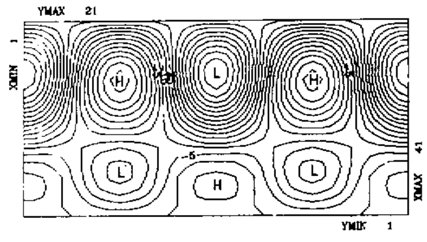

Fig. 3. The stream function field forced by topography alone, nonlinear case, interval: 5, unit:0.1.

Thus, the flow field for the nonlinear stationary wave induced by topography is shown in Fig. 3. It can be seen from Fig. 3 that the high pressure of dipole blocking is very strong, and the low pressure to the south is weaker, which is consistent with the actual observations of dipole blocking. This structure can not be explained by the equivalent modon derived by Mcwilliams (1980). Therefore, it seems that the nonlinear interaction between different stationary modes forced by topography could initiate high–low dipole blockings over the valley regions of topography (i.e. two oceans). And when the height of topography is higher, the strong dipole blocking can be established in two oceans easily. It can be seen from Eq.(12) that the interaction between the topographic waves 2 and 4 would make the energy of the topographic wave 4 transfer into the topographic wave 2. For this reason, the dipole blocking can be produced in the atmosphere.

IV. THE EFFECT OF THERMAL FORCING

In the absence of orographic forcing, i.e., only the thermal forcing is considered in Eq.(4), then, it can be reduced to

$$\frac{\partial}{\partial x}\nabla^2\psi' + \beta_0\frac{\partial\psi'}{\partial x} = -BQ - J(\psi', \nabla^2\psi'), \tag{13}$$

here, by taking the parameter $U \sim 10$ m / s, $L \sim 2200$ km, $f_0 \sim 10^{-4}\mathrm{S}^{-1}$ and $bQ_0 \sim 10^{-11}\mathrm{S}^{-2}$, then, we have $B = 0.44$. Similarly, we may take $\delta = B$ as a small parameter. In this case, the solution of Eq.(13) for ψ' can be expanded asymptotically in the following form

$$\psi' = \delta\psi_1 + \delta^2\psi_2 + \ldots \tag{14}$$

Substitution of Eq.(14) into Eq.(13), we have

$$O(\delta): \frac{\partial}{\partial x}\nabla^2\psi_1 + \beta_0\frac{\partial\psi_1}{\partial x} = -Q, \tag{15}$$

$$O(\delta^2): \frac{\partial}{\partial x}\nabla^2\psi_2 + \beta_0\frac{\partial\psi_2}{\partial x} = -J(\psi_1, \nabla^2\psi_1). \tag{16}$$

In our study, let us assume that the diabatic heating only consists of M mode harmonic waves, that is

$$Q = \frac{1}{2}\sum_{l=1}^{M} Q_l \exp[il(kx - r)]\sin ny + cc, \tag{17}$$

where r denotes the phase shift of the diabatic heating wave relative to the topographic wave, Q_l denotes the amplitude of the l mode diabatic heating, and $Q_l >> Q_{l+1}$ ($l = 1, \cdots M$).

For the sake of simplicity, we may take $M = 2$. In this case, Eq.(17) will be reduced to

$$Q = [Q_1\cos(kx - r) + Q_2\cos 2(kx - r)]\sin ny. \tag{18}$$

Substituting Eq.(18) into Eqs.(15) and (16), we can derive the nonlinear stationary wave solution in the form

$$\psi' = -\delta\left\{\frac{Q_1\sin(kx - r)}{k[\beta_0 - (k^2 + n^2)]} + \frac{Q_2\sin 2(kx - r)}{2k[\beta_0 - (4k^2 + n^2)]}\right\}\sin ny$$
$$+ \frac{9\delta^2 Q_1 Q_2 n \sin 2ny}{8[\beta_0 - (k^2 + n^2)]^2[\beta_0 - (4k^2 + n^2)]}\cos(kx - r). \tag{19}$$

In this section, we shall only discuss the behavior of the nonlinear staionary wave forced by diabatic heating. Now, take $r=0$, $Q_1=-2.0$ and $Q_2=-0.2$, then, the distribution of diabatic heating is shown in Fig. 4.

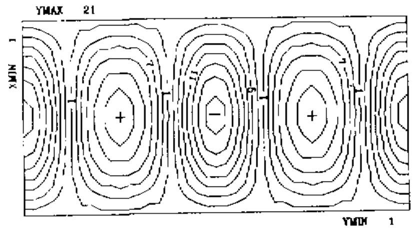

Fig. 4. The distribution of heat source and sink. $r=0$, $Q_1=-2.0$ and $Q_2=-0.2$, interval: 5, unit: 0.1.

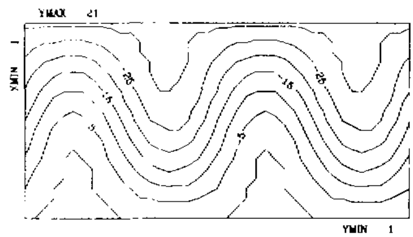

Fig. 5. The stream function field of nonlinear response to the forcing of diabatic heating alone as in Fig. 4. interval 5, unit. 0.1

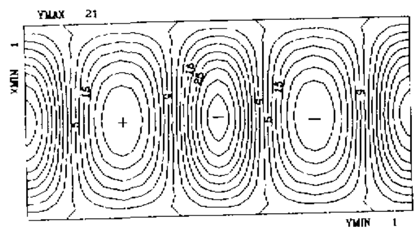

Fig. 6. Same as Fig. 4 but for $r=0$, $Q_1=-4.0$ and $Q_2=-0.4$.

Obviously, in winter, the ocean is a heat source, and the continent is a heat sink. However, the distribution of diabatic heating in summertime is opposite to that in wintertime. Therefore, the flow field for the nonlinear stationary wave excited by diabatic heating is shown in Fig. 5. As can be seen, no blocking occurs in two oceans. If we take $r=0$, $Q_1=-4.0$ and $Q_2=-0.4$, then, the diabatic heating distribution is given in Fig. 6. And the corresponding stream function field of nonlinear stationary wave is shown in Fig. 7. One can see that a blocking high is situated in the west part of heat source and the east part of heat sink. Thus, we may conclude that only when the intensity of diabatic heating is strong enough, can it cause blocking high to form in the atmosphere. Rex (1950) and Sumner (1954) pointed out that the areas where blocking highs appear most frequently in summertime are the eastern parts of the two oceans, which is coincident with the results obtained here in the case of considering only thermal forcing in the summer. Thus, the nonlinear stationary wave solution forced by diabatic heating alone can only expain blocking highs appearing in the summer, but it can neither explain blocking highs nor dipole blockings in the winter. Therefore, the above discussion indicates that the orographic forcing is a main factor for the formation of dipole blocking.

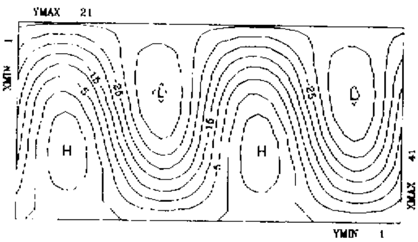

Fig. 7. Same as Fig.5 but for the forcing of diabatic heating alone as in Fig. 6.

V. THE LINEAR STATIONARY WAVE FORCED BY BOTH TOPOGRAPHY AND DIABATIC HEATING

Setting $\delta = B$, Eq.(4) can be rewritten as

$$\frac{\partial}{\partial x}\nabla^2\psi' + \beta_0\frac{\partial\psi'}{\partial x} + F_0\delta\frac{\partial h}{\partial x} = -\delta Q - J(\psi', \nabla^2\psi' + F_0\delta h), \tag{20}$$

where $F_0 = \varepsilon / \delta$, $\varepsilon = 0.66$ and $\delta = 0.44$ and the linearized form of Eq.(20) is

$$\frac{\partial}{\partial x}\nabla^2\psi' + \beta_0\frac{\partial\psi'}{\partial x} + F_0\delta\frac{\partial h}{\partial x} = -\delta Q. \tag{21}$$

Here, the distribution of topography and diabatic heating are taken to be the forms of Eq.(10) and Eq.(18). We can obtain the linear stationary wave forced by both topography and diabatic heating in the following form

$$\psi' = -\delta\left[\frac{h.F_0}{\beta_0 - (k^2+n^2)}\cos kx + \frac{h_2 F_0}{\beta_0 - (4k^2+n^2)}\cos 2kx\right]\sin ny - \delta\left\{\frac{Q_1\sin(kx-r)}{k[\beta_0-(k^2+n^2)]} + \frac{Q_2\sin 2(kx-r)}{2k[\beta_0-(4k^2+n^2)]}\right\}\sin ny. \tag{22}$$

In the following calculation, we take the parameters $h_1 = 1.5$, $h^2 = 0.3$, $r > 0$, $Q_1 < 0$ and $Q_2 < 0$ (winter). The distribution of wintertime diabatic heating for the parameters $r = 0$, $Q_1 = -1.0$ and $Q_2 = -0.1$ is shown in Fig. 8.

Clearly, the heat source is in the centre of the ocean, consequently, the linear blocking highs forced by both topography and diabatic heating are situated in the west regions of the two oceans as shown in Fig. 9. When the heat source is in the east of the ocean (anomalous year) under the condition of taking $r = \frac{\pi}{4}$, $Q_1 = -1.0$ and $Q_2 = -0.1$, the blocking high with respect to the forcings of topography and diabatic heating is located in the centre of the ocean (Figure not shown). When the intensity of diabatic heating is increased, i.e. $r = 0$, $Q_2 = -0.2$ and $Q_1 = -2.0$, then, the linear stationary wave forced by both topography and diabatic heating (see Fig. 4.) is shown in Fig. 10. The blocking high is intensified, and its position is in the further west of the ocean. Similarly, if the heat source is moved to the east of the ocean, then, for $r=\frac{\pi}{4}$, $Q_1=-2.0$ and $Q_2=-0.2$, blocking high will occur slightly to the west of mid-ocean (Figure not shown). It seems that the linear stationary waves forced by both topography and diabatic heating can not explain blocking situations in two oceans in the winter.

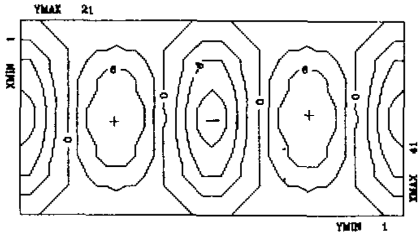

Fig. 8. Same as Fig. 4, but for $r=0$, $Q_1=-1.0$, $Q_2=-0.1$.

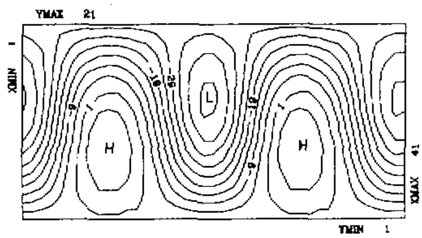

Fig. 9. The stream function field of linear response to the forcings of topography and diabatic heating as in Fig. 1 and Fig. 8. interval: 5, unit: 0.1.

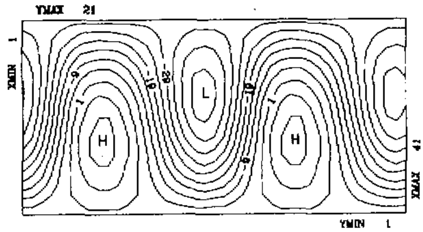

Fig. 10. Same as Fig. 9 but for the forcings of Fig. 1 and Fig. 4.

VI. THE NONLINEAR STATIONARY WAVE FORCED BY BOTH TOPOGRAPHY AND DIABATIC HEATING

In this section, the solution of Eq.(20) for ψ' is expanded into an asymptotic series

$$\psi'=\delta\psi_1+\delta^2\psi_2+\ldots \tag{23}$$

Introducing Eq.(23) into Eq.(20), we get

$$O(\delta):\frac{\partial}{\partial x}\nabla^2\psi_1+\beta_0\frac{\partial\psi_1}{\partial x}+F_0\frac{\partial h}{\partial x}=-Q, \tag{24}$$

$$O(\delta^2):\frac{\partial}{\partial x}\nabla^2\psi_2+\beta_0\frac{\partial\psi_2}{\partial x}=-J(\psi_1,\nabla^2\psi_1+F_0h). \tag{25}$$

Similar to the above discussion, the distributions of topography and diabatic heating are assumed to have the forms of Eq.(10) and Eq.(18), then, we may obtain the nonlinear stationary wave forced by topography and diabatic heating in the form

$$\begin{aligned}
\psi' = & -\delta\left[\frac{h_1F_0\cos kx}{\beta_0-(k^2+n^2)}+\frac{h_2F_0}{\beta_0-(4k^2+n^2)}\cos 2kx\right]\sin ny-\delta\left\{\frac{Q_1\sin(kx-r)}{k[\beta_0-(k^2+n^2)]}\right. \\
& \left.+\frac{Q_2\sin 2(kx-r)}{2k[\beta_0-(4k^2+n^2)]}\right\}\sin ny+\delta^2F_0^2\frac{3}{4}nh_1h_2\sin ny\cos kx \\
& \left\{-\frac{3k^2}{[\beta_0-(k^2+n^2)]^2[\beta_0-(4k^2+n^2)]}+\frac{2}{[\beta_0-(k^2+n^2)]^2}\right. \\
& \left.-\frac{2}{[\beta_0-(k^2+n^2)][\beta_0-(4k^2+n^2)]}\right\}+\frac{9\delta^2Q_1Q_2n\sin 2ny}{8[\beta_0-(k^2+n^2)]^2[\beta_0-(4k^2+n^2)]} \\
& \times\cos(kx-r)-\frac{F_0\delta^2nh_1Q_2}{8}\sin(kx-2r)\left\{\frac{9k}{[\beta_0-(k^2+n^2)]^2[\beta_0-(4k^2+n^2)]}\right. \\
& \left.+\frac{5}{k[\beta_0-(k^2+n^2)][\beta_0-(4k^2+n^2)]}\right\}\sin 2ny-\frac{F_0\delta^2nh_2Q_1}{4}\sin(kx+r) \\
& \times\left\{\frac{9k}{[\beta_0-(k^2+n^2)]^2[\beta_0-(4k^2+n^2)]}\right. \\
& \qquad\left.+\frac{5}{k[\beta_0-(k^2+n^2)][\beta_0-(4k^2+n^2)]}\right\}\sin 2ny.
\end{aligned} \tag{26}$$

1. *Winter Case*

(1) Case $r=0$, i.e., the heat source is in the centre of the ocean. Taking $h_1=1.5$, $h_2=0.3$, $r=0$, $Q_1=-1.0$ and $Q_2=-0.1$, the distribution of diabatic heating is shown in Fig. 8, then, the stream function field for the nonlinear stationary wave forced by both topography and diabatic heating is given in Fig. 11.

A dipole blocking occurs a bit to the west of mid–ocean. If the intensity of diabatic heating increases, then the position of dipole blocking for $r=0$, $Q_1=-2.0$ and $Q_2=-0.2$ will be further west of the ocean (Figure omitted).

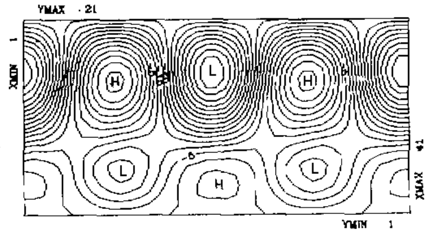

Fig. 11. The stream function field of nonlinear response to the forcings of topography and diabatic heating (i.e. Fig. 1 and Fig. 8), interval: 5, unit: 0.1.

(2) Case $r = -\frac{\pi}{4}$, that is, the heat source is in the west of the ocean. Here, we take $Q_1 = -1.0$ and $Q_2 = -0.1$, then the dipole blocking forced by both topography and diabatic heating is stronger than that in Fig. 11 (see Fig. 12). And it also moves a little eastward, but, it is still located in the western part of the ocean. Therefore, the diabatic heating in the west of the ocean is more favourable for the formation of dipole blocking than in the centre of the ocean. Similarly, if the intensity of diabatic heating is increased, then the dipole blockings formed in two oceans will be intensified and still situated in the west regions of the two oceans.

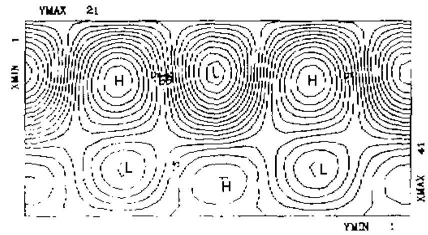

Fig. 12. Same as Fig. 11, but for $r = -\frac{\pi}{4}$, $Q_1 = -1.0$ and $Q_2 = -0.1$.

(3) Case $r = \frac{\pi}{4}$, *in other words, it is equivalent to a heat source in the east of the ocean.* Let us take the parameters $Q_1 = -1.0$ and $Q_2 = -0.1$. In this case, the dipole blocking forced by both topography and diabatic heating is shown in Fig. 13.

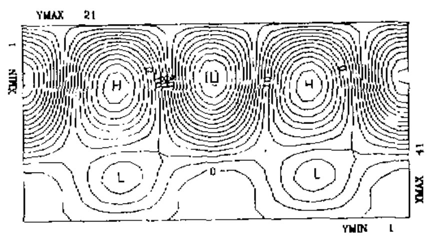

Fig. 13. Same as Fig. 11 but for $r = \frac{\pi}{4}$, $Q_1 = -1.0$ and $Q_2 = -0.1$.

It is clear that the dipole blocking is situated in the west of the ocean, and its intensity is weaker than that in Fig. 11. If we take $Q_1=-2.0$ and $Q_2=-0.2$ again, i.e., a stronger diabatic heating, the strength of dipole blocking is clearly reduced. Thus, the strong diabatic heating source in the east of the ocean is not advantageous to the formation of wintertime dipole blocking.

2. *Summer Case*

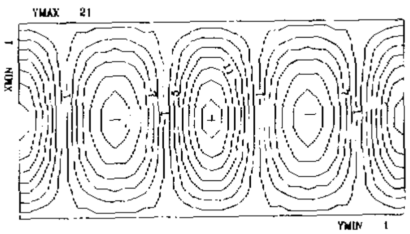

Fig. 14. Same as Fig. 4 but for $r=0$, $Q=2.0$ and $Q_2=0.2$.

Taking $h_1=1.5$, $h_2=0.3$, $r=0$, $Q_1=2.0$, and $Q_2=0.2$, it means that the continent a heat source and the ocean is a heat sink. The distribution of diabatic heating is shown in Fig. 14, and the dipole blocking formed in the oceans is shown in Fig. 15. Obviously, the dipole blocking is situated in the east of the ocean. So the forcings of topography and diabatic heating may initiate dipole blocking in the east of the ocean in the summer.

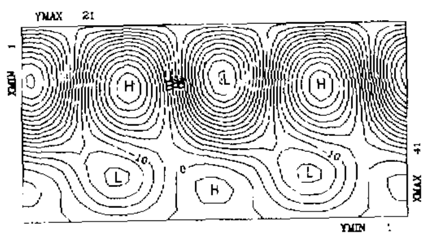

Fig. 15. Same as Fig. 11 but for the forcings of topography (Fig. 1) and diabatic heating (Fig. 14).

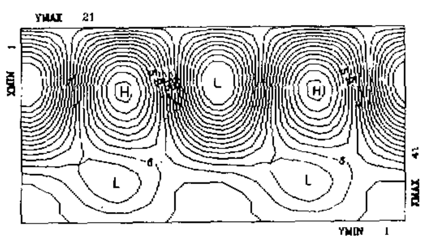

Fig. 16. Same as Fig. 15 but for $r=-\frac{\pi}{4}$, $Q_1=2.0$ and $Q_2=0.2$.

When the diabatic heating moves westward, for simplicity, taking $r=-\frac{\pi}{4}$, $Q_1=2.0$ and $Q_2=0.2$, the result is shown in Fig. 16. It is evident that when the heat source is in the west of the continent, it is not advantageous to the formation of strong dipole blocking in the east of the ocean. However, when the heat source centre is in the east of the continent, the dipole blocking is intensified, see Fig. 17.

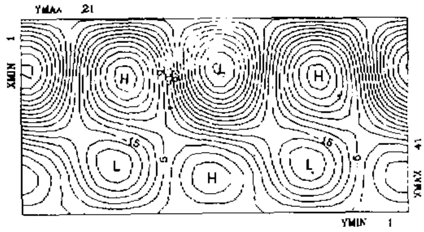

Fig. 17. Same as Fig. 15 but for $r=\frac{\pi}{4}$, $Q_1=2.0$ and $Q_2=0.2$.

Recently, Lejenas and Φkland (1983) carried out a study about Northern Hemisphere blocking during the period of 1950–1979. They found that there are two preferred regions of winter blocking: one is the western part of the Pacific region, another one is the eastern part of the Atlantic region. But in the summer, blocking mainly occurs in the eastern part of the Atlantic region. Thus, the results obtained in our study can partly explain the geographical locations of blocking. Since the observational study of dipole blocking has not been made, it is not clear at this moment that whether the dipole blocking mainly occurs in the west or in the east of the ocean. Nevertheless, it is certain that most of dipole blockings appear in the two oceans. This problem will be studied in another paper. As mentioned above, the forcing of topography is an important factor for the formation of dipole blocking. On the other hand, the location of topographic forcing is basically unchangeable. Thus, the geographical and seasonal variation of dipole blocking mainly depends on the seasonal variation and distribution of diabatic heating. From our studies, we may see that in the winter, dipole blocking mainly occurs in the west of the ocean, but in the summer, the dipole blocking mainly occurs in the east of the ocean. These results are to be tested by some observations in the future.

VII CONCLUSIONS

From above discussion, we may obtain the following conclusions.

(1) When only the forcing of topography is considered in the atmosphere, the nonlinear interaction of the different forced stationary waves may cause dipole blocking in the ocean, but the forcing of diabatic heating alone can not cause dipole blocking. It follows that the forcing of topography is a main factor for the formation of dipole blocking in the atmosphere.

(2) Under the conditions of both topographic forcing and diabatic heating forcing, the nonlinear interaction of the different forced stationary modes may produce dipole blocking. In winter, the dipole blocking mainly occurs in the west of the ocean, and the forcing of diabatic heating in the west of the ocean is advantageous to its formation. In summer, the dipole blocking mainly appears in the east of the ocean, and the heat source in the east of the continent is favourable for the formation of dipole blocking in the east of the ocean.

However, the linear stationary wave forced by both topography and diabatic heating fails to form dipole blocking in the atmosphere.

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