MODULATION OF ATMOSPHERIC THERMAL AND MECHANICAL FORCINGS AND NUMERICAL MODELING OF MEAN MERIDIONAL CIRCULATION

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Received June 21, 1993

ABSTRACT

Upon investigating the relative locations of internal and external forcing and the resultant mean meridional circulation, it was found that thermal forcing and mechanical forcing for the formation of atmospheric mean meridional circulation are modulated by a certain ratio. This ratio is determined by the inherent baroclinity, static stability and absolute vorticity of the atmosphere.

By employing a parameterization scheme for radiative heating and condensation heating, together with the analysis data of the European Center for Medium–range Weather Forecasts, the mean meridional circulation for January was simulated numerically. It was found that latent heat release in the tropics may result in the formation of double–layered Hadley circulation, so do the eddy momentum transfer processes. On the other hand, mean meridional circulations in extra–tropics are mainly determined by external momentum forcing and atmospheric properties of eddy momentum and heat transfer.

Key words: mean meridional circulation, thermal forcing, mechanical forcing

I. INTRODUCTION

As one component of the atmospheric circulation, the mean meridional circulation (MMC) is closely connected with the other components, such as the zonal circulation, the horizontal circulation, and the wind and temperature fields. MMC as a secondary circulation is excited only when the atmospheric hydrostatic and geostrophic balances are destroyed. The atmospheric statues is then modified by the MMC to such extent that a new state of quasi–hydrostatic and geostrophic balances is established (Eady, 1950), Ye and Zhu (1958) indicated that the three–cell structure of the MMC of the atmosphere is formed and maintained by the combining effects of the earth rotation, heterogeneous diabatic heating, eddy transfer, and friction. The analyses of global budgets of angular momentum and sensible heat (Wu and Tibaldi, 1988) showed that the inertia moment of the horizontal branch of the MMC balances the production of angular momentum near the earth's surface and the transfer of angular momentum in the free atmosphere, and the adiabatic heating of its vertical branch balances the diabatic heating and eddy heat transfer in the atmosphere.

Upon analysing the ECMWF annual mean data between 1979 and 1984, Wu and Liu (1987) found that the Hadley circulation at low–latitudes appears as double–cell in the vertical. This phenomenon was confirmed later by the calculation results of Hoskins et al. (1989, see their figures at page 73 and page 105). Because the analyses of tropical mean meridional wind are

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sensitive to the schemes of constructing initial fields (Hollingsworth and Cats, 1981), Trenberth and Olsen (1988) have proposed that the double–layered structure of Hadley circulation was associated with analytic errors. However, the double–layered structure appears only in the winter–hemisphere, and changes regularly from winter to summer (refer to Wu and Liu, 1987, page 20—21), the above phenomenon cannot be explained with analytic errors. Actually, the experiments of Schneider and Linzen (1977) show that the temperature gradient at the earth's surface in tropics can result in direct circulation near the surface, whereas the latent heat release and convection friction can force direct circulation in the upper atmosphere. This means that, the double–layered structure of Hadley circulation can possibly exist. However, their idealized modeling results differ from data analysis considerably, and the cell forced by surface temperature gradient appears only under 800 hPa level. In this study, we tried to investigate the formation and maintenance of the MMC with both numerical analysis and numerical modeling, so as to reveal the probable mechanism for the formation of double–layered Hadley circulation. Firstly, we will discuss the relative location of MMC and its source / sink in Section II. In Section III, we mainly discuss the effect of latent heat release and radiation cooling in forming the Hadley circulation. In Section IV, we will compare the contribution to the maintenance of MMC of internal forcing with that of external forcing. Some conclusions are presented in Section V.

II. LOCATION OF SOURCE / SINK OF MMC

The zonal mean momentum–equation and thermodynamics–equation can be written as

$$[u]_t = \xi[v] - [u]_p[\omega] - (a\cos^2\varphi)^{-1}[u^* v^* \cos^2\varphi]_\varphi - [u^* \omega^*]_p + S, \tag{1}$$

$$[\theta]_t = -a^{-1}[\theta]_\varphi[v] - [\theta]_p[\omega] - (a\cos\varphi)^{-1}[v^* \theta^* \cos\varphi]_\varphi - [\theta^* \omega^*]_p + Q. \tag{2}$$

Here, $\xi = f - (a\cos\varphi)^{-1}[u\cos\varphi]_\varphi$ is zonal mean absolute vorticity. [ ] and * denote zonal mean and the departure from the zonal mean, respectively. The first and second terms on the right of equal–sign are the contributions of MMC to the zonal mean quantities, and the third, fourth and fifth are the contributions to the zonal mean variations of the internal forcing due to horizontal and vertical eddy transfer, and of external forcing, respectively. Express these forcing terms as following:

$$\begin{cases} F_1 = -(a\cos^2\varphi)^{-1} f[u^* v^* \cos^2\varphi]_\varphi = f x_1, \\ F_2 = -f[u^* \omega^*]_p = f x_2, \\ F_3 = fS, \\ F = \sum_{i=1}^{3} F_i = fx = f(x_1 + x_2 + S), \end{cases} \tag{3}$$

$$\begin{cases} H_1 = -\alpha(a\cos\varphi)^{-1}[v^* \theta^* \cos\varphi]_\varphi, \\ H_2 = -\alpha[\omega^* \theta^*]_p, \\ H_3 = \alpha Q, \\ H = \sum_{i=1}^{3} H_i, \end{cases} \tag{4}$$

where $\alpha = (P / P_0)^\kappa R / (ap)$; then define the static stability parameter A, baroclinicity parameter

B and inertia stability parameter C as

$$\begin{cases} A = -(a^2\rho\cos\varphi)^{-1}[\ln\theta]_p, \\ B = (a^2\rho\cos\varphi)^{-1}[\ln\theta]_\varphi, \\ C = (\cos\varphi)^{-1}f\xi, \end{cases} \tag{5}$$

and, according to zonal mean continuity equation:

$$(a\cos\varphi)^{-1}[v\cos\varphi]_\varphi + [\omega]_p = 0, \tag{6}$$

define the stream function ψ via

$$\begin{cases} [v] = (\cos\varphi)^{-1}\psi_p, \\ [\omega] = -(a\cos\varphi)^{-1}\psi_\varphi, \end{cases} \tag{7}$$

finally, by employing zonal mean geostrophic relation

$$[f + u\tan\varphi / a][u] = -a^{-1}[\Phi]_\varphi, \tag{8}$$

static relation

$$[\Phi]_p = -[RT/p], \tag{9}$$

and the definition of equivalent potential temperature

$$[\theta] = [T][P_o / P]^\kappa, \tag{10}$$

we get the thermal wind relation

$$[f + 2u\tan\varphi / a][u]_p = \alpha[\theta]_\varphi. \tag{11}$$

According to Eqs. (3)—(11), the momentum equation (1) and the thermodynamical equation (2) can be written respectively as

$$\begin{cases} f[u]_t = C\psi_p + \delta B\psi_\varphi + F, & (12) \\ \alpha[\theta]_t = -B\psi_p - A\psi_\varphi + H. & (13) \end{cases}$$

Here, $\delta = f/\Gamma = f/[f+2a^{-1}u\tan\varphi] \approx 1$. At stationary state, (12) and (13) compose a simultaneous equation system :

$$\begin{cases} B\psi_\varphi + C\psi_p = -F, \\ A\psi_\varphi + B\psi_p = H, \end{cases} \tag{14}$$

its coefficient determinant is

$$\triangle = B^2 - AC = [\rho g^2]^{-1}([u]_z\Gamma)^2(1 - Rif\xi\Gamma^{-2}), \tag{15}$$

or

$$\triangle = B^2 - AC = [\rho g a^2 \cos^2\varphi]^{-1} f P_E.$$

Here, P_E is the Ertel potential vorticity in P–coordinate system:

$$P_E \equiv -g(f\vec{k} + \nabla\wedge\vec{v}) \cdot \nabla\theta,$$

Ri is Richardson number:

$$Ri \equiv N^2([u]_z)^{-2}.$$

In the earth's atmosphere, $Ri \gg 1$, and $f\xi\Gamma^{-2} \approx 1$, therefore

$$\triangle = B^2 - AC < 0, \tag{16}$$

or equivalently

$$fP_E > 0. \tag{16'}$$

(16)′ means that the atmosphere is symmetrically stable. Then, from (14) , we get solution

$$\begin{cases} \psi_{\varphi} = -a\cos\varphi[\omega] = -\frac{1}{\triangle}(CH+BF), \\ \psi_{p} = a\cos\varphi[v] = \frac{1}{\triangle}(AF+BH). \end{cases} \tag{17}$$

When $F=H=0$, i.e. when (14) is homogeneous, since $\triangle \neq 0$, according to the d'Alembert's principle, there exists only trivial solution, i.e.

$$[v]=[\omega]\equiv 0. \quad (\text{when} F=H=0) \tag{18}$$

This means:

(1) MMC is forced either by external forcing of sources of momentum and heat or internal forcing due to eddy transfer of momentum and heat. Its intensity and location are affected by three parameters, i.e. the atmospheric static stability, inertial stability and baroclinicity. Without forcing or, if internal and external forcings are in balance, MMC cannot exist.

(2) At the center of MMC, $\psi_{\varphi}=\psi_{p}=0$. From (17), we get

$$\begin{cases} BF+CH=0, \\ AF+BH=0. \end{cases} \quad (\text{at the center of MMC}) \tag{19}$$

Now that $\triangle = B^2 - AC \neq 0$, this linear homogeneous equation system possesses only trivial solution

$$F=H=0,$$

i.e. the external sources of momentum and heat are balanced by eddy transfer of momentum and heat, respectively at the center of MMC.

(3) For the location at the lower and upper horizontal branches of MMC where $[\omega]=0$, we obtain from (14)

$$[v]=\frac{H}{B\cos\varphi}=-\frac{x}{\xi}. \qquad (\text{when} \quad [\omega]=0) \tag{20}$$

In the Northern Hemisphere, $\xi>0$, $B<0$, then

$[v]<0$, (when $\chi>0$, or $H>0$)
$[v]>0$; (when $\chi<0$, or $H<0$)

in the Southen Hemisphere, $\xi<0$, $B>0$, then

$[v]>0$, (when $\chi>0$, or $H>0$)
$[v]<0$. (when $\chi<0$, or $H<0$)

This is to say, the region of momentum source (sink) and / or heat source (sink) corresponds to horizontal equatorward (poleward) motion.

(4) For the locations of the pure rising and sinking branches where $[v]=0$, we obtain from (14)

$$[\omega]=-\frac{1}{a\cos\varphi}\psi_{\varphi}=\frac{1}{a\cos\varphi}\frac{fx}{B}=-\frac{1}{a\cos\varphi}\frac{H}{A}. \quad (\text{when} \quad [v]=0)$$

Since $f/B<0$, in the two hemispheres, then

$-[\omega]>0$, (when $\chi>0$, or $H>0$)
$-[\omega]<0$. (when $\chi<0$, or $H<0$)

This then implies that the region of momentum source (sink) corresponds to rising (sinking) motion.

As a summary, the correlation between the location of source / sink of zonal mean momentum and / or heat and the location of MMC can be expressed schematically in Fig. 1a 'and described as follows:

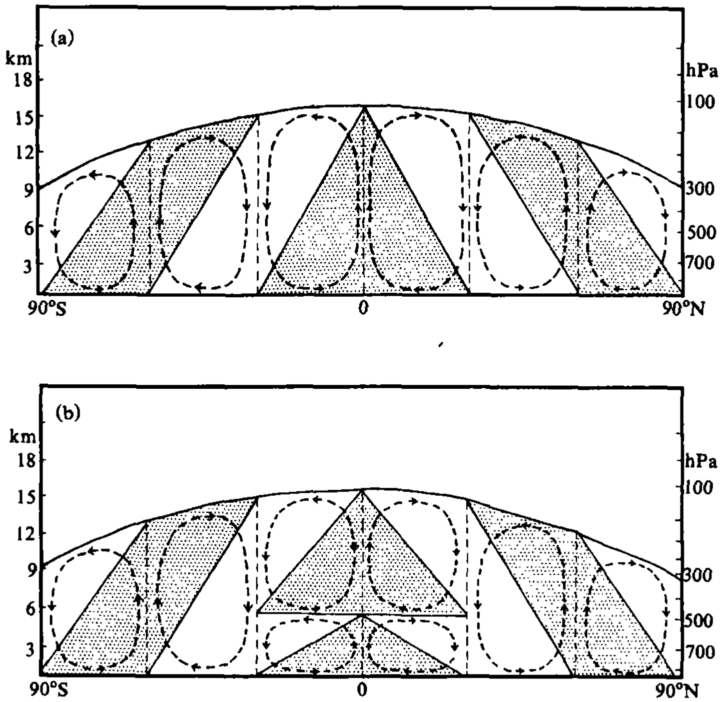

Fig. 1. MMC and the locations of the associated forcing source (hatched area) and sink (unhatched area) of momentum and heat: (a) Typical three–cell circulation; and (b) tropical double–layer Hadley circulation.

For a direct circulation, its rising branch and the near surface branch correspond to the source of momentum and / or heat, and the sinking branch and near tropopause branch correspond to sink. For an indirect MMC, its rising branch and near tropapause branch correrpond to source of momentum and / or heat, and its sinking branch and near surface branch correspond to sink. When there exists double–layered Hadley circulation in the tropics, the distribution of source and sink should appear as in Fig. 1b. In such circumstances there must be double–layered source and sink of momentum and heat.

It is worthwhile to note that definite proportional relations exist between thermal forcing and mechanical forcing to MMC at the spots where Eqs. (20) and (21) are satisfied; and these proportions are modulated by the internal baroclinicity, static stability and inertia stability of the atmosphere. In other words, the dynamic forcing and the thermal forcing of the MMC are modulated according to the following relations.

At the spots of $[\omega]=0$, the ratio of momentum source to heat source is proportional to the ratio of absolute vorticity to baroclinicity, and at the spots of $[v]=0$, the ratio of the product of momentum source and f to the heat source is proportional to the ratio of baroclinicity to static stability, or proportional to the tangent of the slantwise between the isentropic surface and the isobaric surface.

III. THERMAL FORCING AND HADLEY CIRCULATION

By using thermal wind relation (11), the equation of MMC can be obtained from momentum equation (12) and thermodynamics equation (13) as

$$(A\psi_\varphi)_\varphi + 2B\psi_{\varphi p} + (C\psi_p)_p + B_p\psi_\varphi + B_\varphi\psi_p$$
$$= (F_1 + F_2 + F_3)_p + (H_1 + H_2 + H_3)_\varphi = F_p + H_\varphi, \quad (22)$$

the approximate relation $|[u]| / (a\Omega\cos\varphi) \ll 1$ has been employed in obtaining the above equation. The nature of Eq. (22) is determined by the sign of $\triangle = B^2 - AC$ (or the magnitude of Richardson number). Noticing that the atmosphere is symmetrically stable ($fP_E > 0$) in general, from (16) or (16)′, we know that Eq. (22) is an elliptic equation. Since homogeneous elliptic equation has no extremes at any inner point, and since the boundary condition of the global integration is approximately zero, from (22), when $F = H \equiv 0$, then $\psi \equiv 0$. This is to say, if there is no forcing, or if internal source and external source are offset each other everywhere, MMC cannot exist in the atmosphere. Besides, since Eq. (22) is linear, MMC can be considered as the sum of circulations due to all individual forcings.

To get the numerical solution of Eq. (22), in vertical direction the atmosphere is divided into 40 layers with an interval of 25 hPa, and in horizontal direction into 60 belts with a grid interval of 3 degrees from the south pole to the north pole. Now, there are 61 × 41 grid points. At the two polar regions and at $P = 0$, the boundary value of ψ is set to zero. At $P = 1000$ hPa, we get the lower boundary value of ψ by using the scheme of central finite difference after obtaining ψ_φ from Eq. (17). In this study, our main interest is the MMC in January. The grid values of $[u]$, $[T]$, $[u^* v^*]$, $[u^* \omega^*]$, $[v^* T^*]$, $[\omega^* T^*]$, external momentum source F_3, and heat source H_3 can be interpolated from the data provided in Wu and Liu (1987). These were calculated from the 13–layer and 5–year (Sep., 1979—Aug., 1984) monthly mean January ECMWF data. With these data, Eq. (22) can be solved by super–relaxation to get the distribution of MMC.

Without considering other factors, and only substituting the aforementioned diabatic heating H_3 into (22), we obtain the MMC as shown in Fig. 2. The effect of diabatic heating is weaker at extratropical regions. On the other hand, it excites strong direct circulations in low–latitude regions, whose centers are at 250 hPa and 800 hPa with intensities 14×10^3 and 12×10^3 kg s^{-3}, respectively. Because circulations excited by sensible heating are weak and lie under 900 hPa (Schneider and Linzen, 1977), one can refer that these tropical circulations are mainly forced by latent heat release and radiation heating.

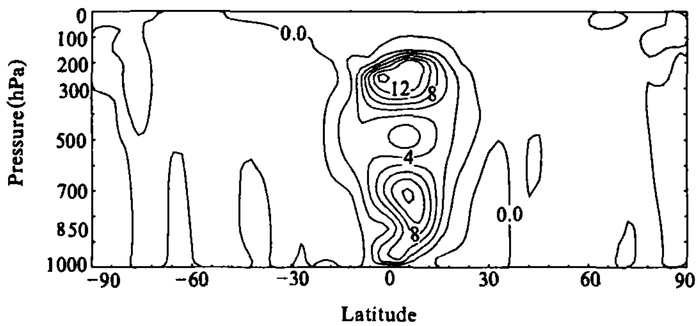

Fig. 2. MMC excited by atmospheric external heat source H_3 in January. The contour interval is 2×10^3 kg s^{-3}. The data of external heat source H_3 came from Wu and Liu which were calculated from thermodynamics equation and from ECMWF 5–year data.

In order to study the impact of radiation heating and tropical convection on MMC, we introduce the following parameterization scheme of radiation heating:

$$Q_{rad} = \left(\frac{P_0}{P}\right)^{\kappa} \frac{T_e(\varphi,p) - [T(\varphi,p)]}{\tau(p)}. \tag{23}$$

Here, $[T(\varphi, p)]$ is the distribution of zonal mean temperature in January. $\tau(p)$ is the time scale of radiation heating in P–coordinate system:

$$\tau(p) = \begin{cases} 30 \text{ days}, & (P \geqslant 300 \text{ hPa}) \\ (0.01P - 1) \times 15 \text{ days}, & (200 \text{ hPa} \leqslant P < 300 \text{ hPa}) \\ 15 \ days, & (P < 200 \text{ hPa}) \end{cases}$$

$T_e(\varphi,p)$ is the radiation equilibrium temperature:

$$T_e(\varphi,p) = 0.5\left\{T(\varphi) + \Gamma H_0 \ln\frac{P}{P_0} + \left|T(\varphi) + \Gamma H_0 \ln\frac{P}{P_0}\right|\right\} + 213.$$

Here, Γ is dry adiabatic lapse rate, assumed as constant; H_0 is the height of standard atmosphere; $T(\varphi)$+213 is idealized radiational equilibrium temperature at 1000 hPa. Supposing that $T(\varphi)$ is maximum at 15°S in January, then we let

$$T(\varphi) = 25\left\{1 - 2\sin^2[0.86(\varphi + 15)]\right\} + 60.$$

The above radiation equilibrium temperature (T_e) gives a distribution of constant skin temperature at the upper limit of the atmosphere to ensure the zonal wind being finite there. Eq. (23) indicates that radiation cooling is proportional to the departure of temperature from the radiation equilibrium state. Rodgers and Walshaw (1966) proved that the radiation scheme described with Eq. (23) is quite precise for the earth's atmosphere.

The following is the parameterization of condensation heating:

$$Q_{con} = \frac{1}{C_p}\left(\frac{P_0}{P}\right)^{\kappa} \tilde{Q}(p) \exp\left[-\left(\frac{\varphi - \varphi_0}{d\varphi}\right)^2\right]. \tag{24}$$

Here, φ_0 is the latitude of condensation heating center, $d\varphi$ presents the Gaussian half–width of the tropical rainfall profile. $\tilde{Q}(p)$ imitates the vertical structure of condensation heating:

$$\tilde{Q}(p) = \begin{cases} C_Q(P - P_t)/(P_1 - P_t), & p_t \leqslant P < P_1 \\ C_Q, & P_1 \leqslant P < P_2 \\ C_Q(P_b - P)/(P_b - P_2), & P_2 \leqslant P < P_b \end{cases} \tag{25}$$

and it satisfies

$$\int_{P_t}^{P_b} \tilde{Q}(p) \frac{dp}{g} = L\rho_{water} P_r. \tag{26}$$

Here, P_t and P_b are the heights of top and bottom of condensation layer, respectively, and P_1 and P_2 are the heights of top and bottom of uniform condensation layer, respectively. L is latent heat; P_r is precipitation rate; ρ_{water}, the density of water.

The parameterization of convection friction is in the simple flux form.

$$F_{con} = -g\frac{\partial}{\partial P}[M_c(u - u_c)]. \tag{27}$$

Here, M_c is mass flux of cumulus, and it can be obtained from the relation

$$Q_{\text{con}} = \omega_c \frac{\partial \theta}{\partial P} = -gM_c \frac{\partial \theta}{\partial P}, \tag{28}$$

if Q_{con} is known.

C_Q could be calculated from (25) and (26) by using observed annual precipitation rate, and the distribution of tropical precipitation could be simulated by adjusting $d\varphi$. Supposing that the latent heat release is realized by both shallow convection and deep convection, and the ratio is 8:7, we can adapt the parameters listed in Table 1.

Table 1. Parameters for Deep and Shallow Convections

	P_t (hPa)	P_1 (hPa)	P_2 (hPa)	P_b (hPa)	P_r (m / year)
Deep Convection	100	200	300	500	0.8
Shallow Convection	500	800	850	900	0.7

To model winter situation, take $\varphi_0 = 6°$S, $d\varphi = 3°$ in Eq. (24). By substituting the above parameters into (23) and (24), the distribution of diabatic heating can be obtained, and the simulated MMC is shown in Fig. 3. Cumulus convection forces direct meridional circulation on the south and north sides of the heating area (Fig.3a). However, the circulation on the south side is weaker and the center is near 250 hPa; whereas that on the north side is stronger and the forced near–equator circulation centers are at 250 hPa and 850 hPa with intensities of 13×10^3 and 11×10^3 kg s^{-3}, respectively. The center of the tropical direct circulation excited by the differential radiation heating between high–latitude and low–latitude is at about 500 hPa, but the intensity is only one third of the circulation forced by condensation heating (figure omitted). Consequently, the meridional circulation forced jointly by the cumulus and radiation heating (Fig. 3b) is similar to Fig. 3a. Comparing Fig. 3 with Fig. 2, one can conclude that the heating rate described with the parameterization scheme in Eqs. (23) and (24) can model the zonal mean heating of the atmosphere quite well, and the MMCs forced by them are very similar to each other.

IV. SIMULATING MMC

The MMC forced by radiation and condensation heating cannot produce enough heat transfer to balance the heating difference between high–latitude and low–latitude, and this should cause too large temperature difference between north and south and a too strong

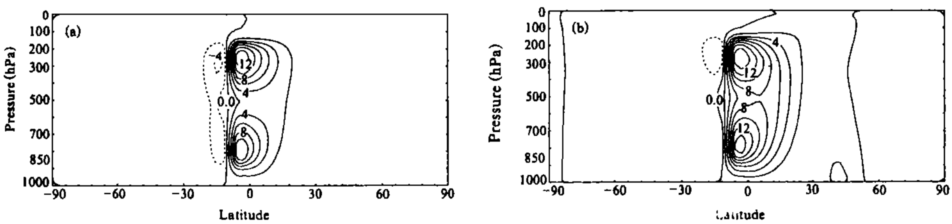

Fig. 3. January MMC modeled by employing parameterization scheme of condensation heating (24) and that of radiation heating (23). The contour interval is 2×10^3 kg s^{-3}: (a) MMC excited by condensation heating; and (b) MMC excited jointly by condensation and radiation heating.

westerly jet, thus resulting in the development of disturbance of barotropic and baroclinic instability. In addition, topography and the thermal contrast between ocean and land can also excite atmospheric disturbances. The development of disturbance can efficiently reduce the temperature difference between north and south. The eddy transfer of momentum and heat can destroy the atmospheric geostrophic and hydrostatic balances. Therefore, ageostrophic secondary circulation is excited so as to establish a new atmospheric balance state (Eady, 1950; Green, 1970). To analyse the forcing effect of eddy transfer on MMC, we introduce (F_1+F_2) and (H_1+H_2) from Wu and Liu (1987) into Eq. (22). The calculated MMC is shown in Fig. 4. The internal momentum forcing and thermal forcing respectively excite three–cell circulation in both the Northern and Southern Hemispheres. In general, the center of the meridional circulation excited by eddy momentum forcing is at 500 hPa, and the intensity of the indirect meridional circulation in the mid–latitudes is 1—3 times more than that of the direct circulation in the tropics. It is noticeable that Hadley circulation appears as a single complete cell in the summer hemisphere, whereas in winter hemisphere there is a center at 250 hPa and another at 650 hPa both with intensity of 2×10^3 kg s^{-3}. Since the main difference in eddy transfer between the Northern and Southern Hemispheres in January is the planetary scale eddy transfer, we deduce that the existence of the upper–layer Hadley circulation in winter hemisphere may be partly due to the development of planetary scale eddies in winter half–year and the associated angular momentum transfer. The latitude location of the MMC center forced by eddy heat transfer in the Southern Hemisphere coincides with that forced by eddy momentum transfer, whereas in the Northern Hemisphere the former moves about 10 degrees northward than the latter.

Analyses show that the MMC excited by momentum forcing of cumulus friction is weaker (figure not shown). Now, we employ the momentum source F_3 from Wu and Liu (1987) to study its contribution to the formation of MMC. Figure 5 demonstrates the MMC forced by F_3. The centers of MMC mostly concentrate in the lower troposphere. Except that the centers of the southern polar circulation and of the northern Ferrel circulation appear at 750 hPa and 800 hPa, respectively, the other centers all appear at 900 hPa. The intensity of the lower–layer Hadley circulation in the Northern Hemisphere retains as high as 15×10^3 kg s^{-3}. According to the discussion in Section II, the external momentum sources which excite the MMC in Fig. 5 should be mainly concentrated near the earth's surface, and they have good correspondence

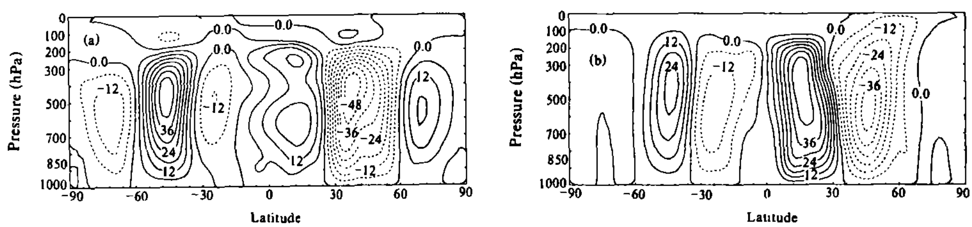

Fig. 4. MMC forced by atmospheric eddy transfer in January (contour interval is 6×10^2 kg s^{-3}): (a) MMC forced by horizontal and vertical eddy momentum transfer; and (b) MMC forced by horizontal and vertical eddy heat transfer.

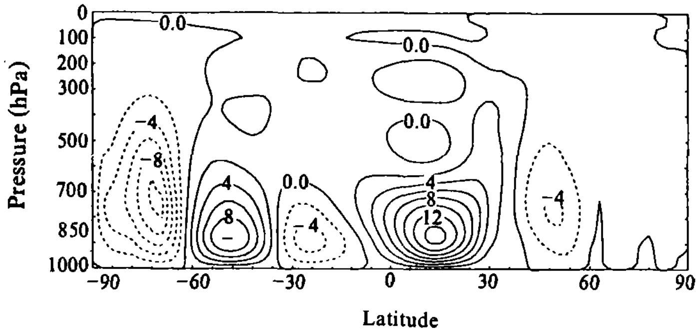

Fig. 5. MMC excited by atmospheric external momentum source (F_3) in January. The contour interval is 2×10^3 kg s^{-3}. External momentum source F_3 was calculated from the u–momentum equation and ECMWF data (Wu and Liu, 1987).

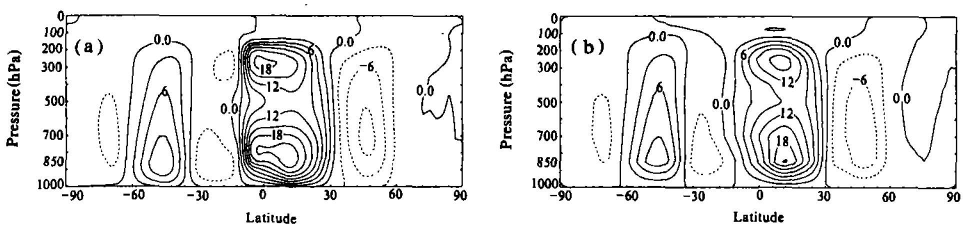

Fig. 6. Distribution of MMC in January. The contour interval is 3×10^3 kg s^{-3}. (a) MMC simulated by employing the parameterization schemes of radiation and condensation heating and other internal–external forcing. Refer to text for details; and (b) MMC calculated directly from zonal mean wind [ν] and [ω] adopted from Wu and Liu, (1987).

with the distribution of surface zonal wind belts. This means that, friction moment is the main factor which excites the circulation near the earth's surface. It is also interesting to note that there is a weak direct circulation near 300 hPa in the tropical region of the Northern Hemisphere. Its intensity is about 3×10^3 kg s^{-3}, slightly stronger than the MMC excited by internal momentum forcing. In addition, an indirect circulation is found between the two direct circulations. It seems that the formation of the upper–layer Hadley circulation in the tropics may be due to the existence of momentum source in the upper troposphere.

Briefly speaking, the center of the upper–layer Hadley circulation near 250 hPa is mainly forced by diabatic heating, and also affected by external momentum forcing and eddy momentum transfer. The ratio of the contribution among these three factors is 15:3:2.

If the diabatic heating scheme described in (23) and (24) and the forcing factors (H_1, H_2 and F) calculated from data and described in this section are added together and substituted into (22), the excited MMC is shown in Fig.6a. There are typical three cells in each hemisphere, and very similar to the MMC (Fig.6b, adopted from Wu and Liu, 1987) calculated directly from the observed January [v] and [ω]. In both simulation and analyses, the Hadley circulation of the Northern Hemisphere possesses obvious double–layer structure. The simulation results also show that to the formation of the lower Hadley cell the contributions of dynamic forcing and thermal forcing are comparable.

V. CONCLUSIONS

MMC is excited by atmospheric external momentum and / or heat forcing and internal transfer processes. There exist definite proportion between zonal mean momentum forcing (F) and heat forcing (H). At points of $[\omega]=0$ (i.e. vertical motion is zero), the ratio of dynamic forcing to thermal forcing is proportional to the ratio of absolute vorticity to baroclinicity. At points of $[v]=0$ (i.e. horizontal motion is zero), the ratio of the product of dynamic forcing and Coriolis parameter to the thermal forcing is proportional to the ratio of baroclinicity to static stability.

The latent heat release from tropical deep convection can produce heat source in the upper troposphere and result in upper–layer Hadley circulation. External momentum source and eddy transfer have contribution to the formation of upper–layer Hadley circulation as well, but their effect is weaker than that of the former. In the formation of the lower–layer Hadley circulation, the contributions of thermal forcing and dynamic forcing are comparable. In the formation of the indirect circulation in the mid–latitudes, the effect of diabatic heating is weaker, whereas external momentum forcing and eddy transfer of momentum and heat are both important.

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