Eddy Heat Flux Convergence in the Troposphere and Its Effect on the Meridional Circulation and Ozone Distribution

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ABSTRACT

In this study the vertical convergence of the eddy heat flux, found as a forcing term in the thermodynamic energy equation of the transformed Eulerian mean formulation, is estimated in the troposphere and in the lower stratosphere from climatalogical data. Results show that while the heating rates caused by these eddy effects are small in the stratosphere they may play an important role in tropospheric circulation. The eddy-caused additions to the forcing field are seen as a region of significant cooling in the midlatitudes at the midtroposphere level and of weak heating throughout the tropical region. This net global cooling is important in balancing net global heating. In addition, the heating due to meridional heat flux is found to dominate compared to heating due to the vertical heat flux. To study circulation changes, the residual mean circulation is calculated with and without the estimated eddy heating effects. The added forcing causes additional circulation in each hemisphere that coincides with the primary circulation due to zonal-mean diabatic heating. Therefore, the eddy heat flux convergence has a significant role in enhancing the zonal-mean residual circulation in the troposphere.

1. Introduction

A major concern to scientists and policy makers is the effect of human-related emissions of chemical species in the atmosphere and their effect on ozone in the stratosphere. The chlorofluorocarbons released on the earth surface, for example, have long chemical lifetimes in the troposphere and, hence, are eventually transported to the stratosphere where their destruction occurs. It is essential to understand the transport processes of such chemical species in the troposphere as well as in the stratosphere to estimate the mass flow between the lower and middle atmospheres. Therefore, a consistent treatment of the transport mechanisms throughout the troposphere and stratosphere is very important.

The modeling tool that has been widely used recently to investigate this ozone problem is the zonally averaged two-dimensional chemical transport model. Among the two-dimensional models, the most popular formulation for the zonal averaging is the transformed Eulerian mean (TEM) equations. The residual mean meridional circulation (RMMC) of the TEM system can be obtained by solving the thermodynamic energy equation and the continuity equation simultaneously. A term in the thermodynamic energy equation usually neglected in solving for the RMMC is the vertical convergence of the eddy heat flux, which does not appear in the equation when scaled quasigeostrophically (Dunkerton et al. 1981; Andrews et al. 1987). In general, the eddy heat flux convergence is much smaller than the zonal-mean heating in the stratosphere (Hitchman and Leovy 1986; Shine 1989), but even when using accurate radiative transfer schemes for the zonal-mean heating in the middle atmosphere, models still use a simple treatment of diabatic heating for the circulation in the troposphere, and the eddy heat flux convergence is neglected. Although the eddy heat flux convergence is negligible in the stratosphere, it may be significant in the troposphere. Furthermore, in the mesosphere the eddy heat flux convergence may be important since gravity wave breaking leads to a significant downward heat flux (Coy and Fritts 1988).

Our main concern in this study will center on the zonal-mean circulation for chemical transport in the troposphere. The eddy heat flux consists of two terms, the meridional heat flux multiplied by the isentropic slope and the vertical heat flux, as shown in the next section. The role of the isentropic slope seems to be important when we consider the maximum slope of the isentropic surfaces is much larger in the troposphere than in the stratosphere. We are interested in estimating the eddy heat flux convergence mainly because of its possible effect on the RMMC. Another motivation for estimating the eddy heat flux convergence is that this

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term might contribute to the global equilibrium of the diabatic heating, even though in the stratosphere, as noted by Shine (1989), the eddy effect is too small to change the net radiative heating or cooling.

In this study we estimate the magnitude of heating in the thermodynamic energy equation due to the vertical convergence of the eddy heat flux by using long-time climatological data. We compare the effects of meridional and vertical eddies and also compare total eddy effects with the zonal-mean diabatic heating produced by the Lawrence Livermore National Laboratory (LLNL) two-dimensional model. We estimate the difference of the meridional circulation caused by the eddy heat flux convergence in the troposphere. Finally, we evaluated the effect of the eddy heat flux convergence on the distribution of column ozone using the LLNL 2D model. As our model results will show, the addition of tropospheric eddy heat flux provides an total ozone distribution in better agreement to observations when compared to the ozone distribution without the eddy term.

2. Formulation

In the log pressure coordinate system the thermodynamic energy equation in the TEM system is written as (Andrews et al. 1987)

$$\frac{\partial\bar{\theta}}{\partial t} + \bar{v}^* \frac{\partial\bar{\theta}}{\partial y} + \bar{w}^* \frac{\partial\bar{\theta}}{\partial z} = \bar{Q} + \bar{E}, \tag{1}$$

where the eddy heat flux convergence $\bar{E}$ is defined by

$$\bar{E} = -\frac{1}{\rho_0}\frac{\partial}{\partial z}\left[\rho_0\left(\overline{v'\theta'}\,\frac{\partial\bar{\theta}/\partial y}{\partial\bar{\theta}/\partial z} + \overline{w'\theta'}\right)\right] \tag{2}$$

and primes denote departures from zonal mean.

Equation (2) can be rewritten as

$$\bar{E} = -\frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\frac{\rho_0\overline{\mathbf{v}'\theta'}\cdot\nabla\bar{\theta}}{\partial\bar{\theta}/\partial z}\right), \tag{3}$$

where $\mathbf{v}' = (v', w')$. The above relation shows that the term within the square brackets of (2) is proportional to the component of eddy heat flux parallel to the mean potential temperature gradient (Dunkerton et al. 1981; Holton 1981). Thus, there is a close relationship between $\bar{E}$ and the isentropic slope. It also can be shown that for linear waves in the absence of transience and diabatic effects the eddy forcing in (3) vanishes (Andrews and McIntyre 1976, 1978; Plumb 1979; Holton 1980, 1981). The wave activity in the middle atmosphere is thought to be nearly steady and conservative; the eddy effects are thus usually ignored in the calculation of the residual circulation (e.g., Dunkerton 1978; Hitchman and Leovy 1986; Solomon et al. 1986; Wuebbles et al. 1987; Gille et al. 1987; Rosenfield et al. 1987; Johnston et al. 1989; Shine 1989).

The situation for transient and/or damped waves is discussed in detail in the work by Plumb (1979). Especially for the steady and nonconservative wave, the eddy heat flux becomes

$$\bar{E} = -\frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0\frac{\overline{Q'\theta'}}{\partial\bar{\theta}/\partial z}\right), \tag{4}$$

or by representing Q' by a linearized Newtonian cooling, then

$$\bar{E} = \frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0\frac{\alpha\overline{\theta'^2}}{\partial\bar{\theta}/\partial z}\right). \tag{5}$$

Shine (1989) noted that $\bar{E}$ calculated by using (5) from the linearized cooling coefficients derived from the radiation scheme and the Barnett and Corney (1985) temperatures does not exceed 0.2 K day^{-1} anywhere in the middle atmosphere. Hitchman and Leovy (1986) also estimated $\bar{E}$ by wave temperatures from *Nimbus 7* Limb Interferometer Monitor of the Stratosphere observation and found that for most situations it was much smaller than the other terms in the equation. By the above argument and estimation, neglecting $\bar{E}$ seems to be justified in the middle atmosphere.

Although the eddy heat flux convergence term is usually neglected, its effect is not always ignored in the middle atmosphere. Instead of completely neglecting the eddy heat flux convergence, Garcia and Solomon (1983) and Hitchman and Brasseur (1988) parameterized the heat fluxes through the product of the diffusion coefficients and the mean gradients. The heat fluxes are assumed that $\overline{v'\theta'} = -K_{yy}\partial\bar{\theta}/\partial y$ and $\overline{w'\theta'} = -K_{zz}\partial\bar{\theta}/\partial z$ In that case, (2) becomes

$$\bar{E} = \frac{1}{\rho_0}\frac{\partial}{\partial z}\left\{\rho_0\left[K_{yy}\left(\frac{\partial\bar{\theta}/\partial y}{\partial\bar{\theta}/\partial z}\right)^2 + K_{zz}\right]\frac{\partial\bar{\theta}}{\partial z}\right\}. \tag{6}$$

In determining K_{yy} and K_{zz}, they assume K_{yy} is due entirely to Rossby wave absorption and K_{zz} is due entirely to gravity wave absorption. In a recent model of Brasseur et al. (1990) the contribution of eddies is assumed to be related to only vertical eddy exchanges, so that $\bar{E}$ is parameterized, without meridional component, by

$$\bar{E} = \frac{1}{\rho_0}\frac{\partial}{\partial z}\left(\rho_0 K'_{zz}\frac{\partial\bar{\theta}}{\partial z}\right). \tag{7}$$

In the upper mesosphere, where the gravity wave breaking is essential in the momentum balance, (7) may be a sufficient parameterization. In other regions, however, the relative importance of the horizontal and vertical eddy effects has not been evaluated.

In the troposphere the wave activity may be different from that of the middle atmosphere and may not satisfy the steady and conservative conditions. In fact, in the troposphere the transient eddies are as important as standing eddies, as shown by long-term climatology (e.g., Oort 1983). Also, due to the other heating sources such as the latent heating and sensible heating, the longitudinally dependent heating may significantly contribute to the eddy effects. A convenient variable signifying the eddy effects in the troposphere is the slope of the isentropic surfaces in the meridional plane. In terms of this slope (2) may be rewritten as

$$\bar{E} = \frac{1}{\rho_0}\frac{\partial}{\partial z}[\rho_0(\overline{v'\theta'}\bar{S} - \overline{w'\theta'})], \quad (8)$$

where the isentropic slope $\bar{S}$ is defined by

$$\bar{S} = -\frac{\partial\bar{\theta}/\partial y}{\partial\bar{\theta}/\partial z}. \quad (9)$$

The isentropic slopes are significantly larger in the troposphere than in the middle atmosphere, especially in the middle-latitude region (see Fig. 1). The order of magnitude of typical $\bar{S}$ is 10^{-3}, and the order of magnitude of typical $\bar{v}^*$ and $\bar{w}^*$ are 10.0 and 0.01. Therefore, it is difficult to determine whether there is a dominant eddy effect in (10) between the meridional and vertical direction without calculations using real data.

It is also important in studies of the troposphere, especially in two-dimensional models, to mention the importance of the latent heating. As will be discussed later, the simulations in this paper use the LLNL 2D chemical transport model where latent heating is parameterized and proportional to observed latitude-dependent monthly rainfall data (see Grant et al. 1987 for details). These latent heating rates are always difficult to estimate and do provide some level of uncertainty. However, in this study we investigate the differences seen in the circulation pattern via the inclusion of the eddy heating term and not the circulation pattern itself. Hence, while the absolute accuracy of the latent heating term plays a major role in the circulation pattern, it plays a secondary role in the study.

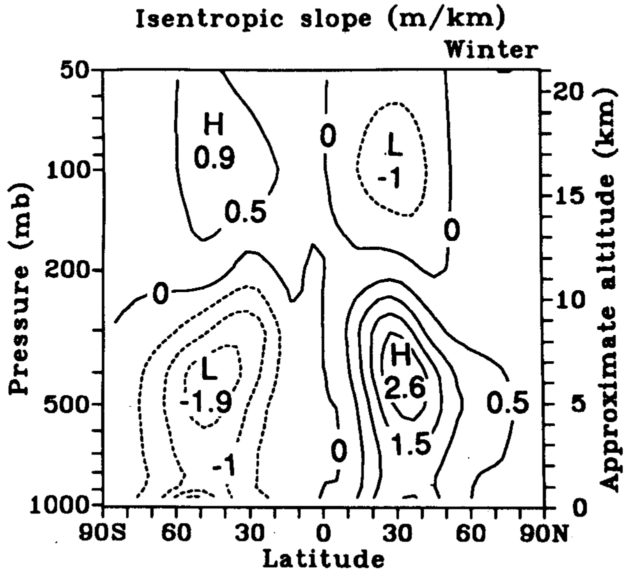

FIG. 1. Slope of the isentropic surface (m km^{-1}) in winter (average of December, January, and February) based on observed temperature (Oort 1983). Contour interval is 0.5. Dotted lines denote negative values.

3. Estimation of eddy heating rates from the climatological data

From climatological data we can estimate temporal- and zonal-averaged eddy effects. To use both the zonal and the temporal mean, it is convenient to use different notations from what we used in the preceding section. In the following the zonal mean and its perturbation are denoted by a square bracket and an asterisk, while the monthly mean and its perturbation are denoted by an overbar and a prime. An arbitrary variable may be written as

$$A = [A] + A^* = \bar{A} + A'. \quad (10)$$

By using the above notations, the eddy heat flux convergence in the form of (8) can be written, after being averaged zonally first and temporally second, as

$$\overline{[E]} = \frac{1}{\rho_0}\frac{\partial}{\partial z}\{\rho_0(\overline{[v^*\theta^*]}\overline{[S]} - \overline{[w^*\theta^*]})\}, \quad (11)$$

where the transient fluctuation of the zonal-mean variables $\overline{[v^*\theta^*]'[S]'}$ was neglected. The flux terms in the right-hand side of (11) can be written as

$$\overline{[v^*\theta^*]} = [\bar{v}^*\bar{\theta}^*] + [\overline{v'\theta'}], \quad (12a)$$

$$\overline{[w^*\theta^*]} = [\bar{w}^*\bar{\theta}^*], \quad (12b)$$

where the heat transport by transient zonal-mean circulation $\overline{[v]'[S]'}$ is neglected. In the vertical direction heat transport by both transient eddies and transient zonal-mean circulation are neglected.

Using the definitions of potential temperature and the vertical velocity $\omega = dp/dt$, (12) is rewritten as

$$\overline{[v^*\theta^*]} = (p_s/p)^\kappa\{[\bar{v}^*\bar{T}^*] + [\overline{v'T'}]\}, \quad (13a)$$

$$\overline{[w^*\theta^*]} = -\frac{H}{p}(p_s/p)^\kappa[\bar{\omega}^*\bar{T}^*]. \quad (13b)$$

The eddy heat flux convergence $\bar{E}$ defined by (2) is represented in terms of the rate of change of potential temperature. For the sake of convenience of presentation the heating rate e and its meridional and vertical contributions e_y and e_z are defined, in terms of temperature change, as

$$E \equiv e(p_s/p)^\kappa \equiv (e_y + e_z)(p_s/p)^\kappa, \quad (14)$$

where the overbar and square brackets are omitted for simplicity. We will call e the "eddy heating" hereafter. Substituting (13) into (11) and using (14) gives

$$e_y = \left(\frac{\partial}{\partial z} + \frac{\kappa - 1}{H}\right)ST_v, \quad (15a)$$

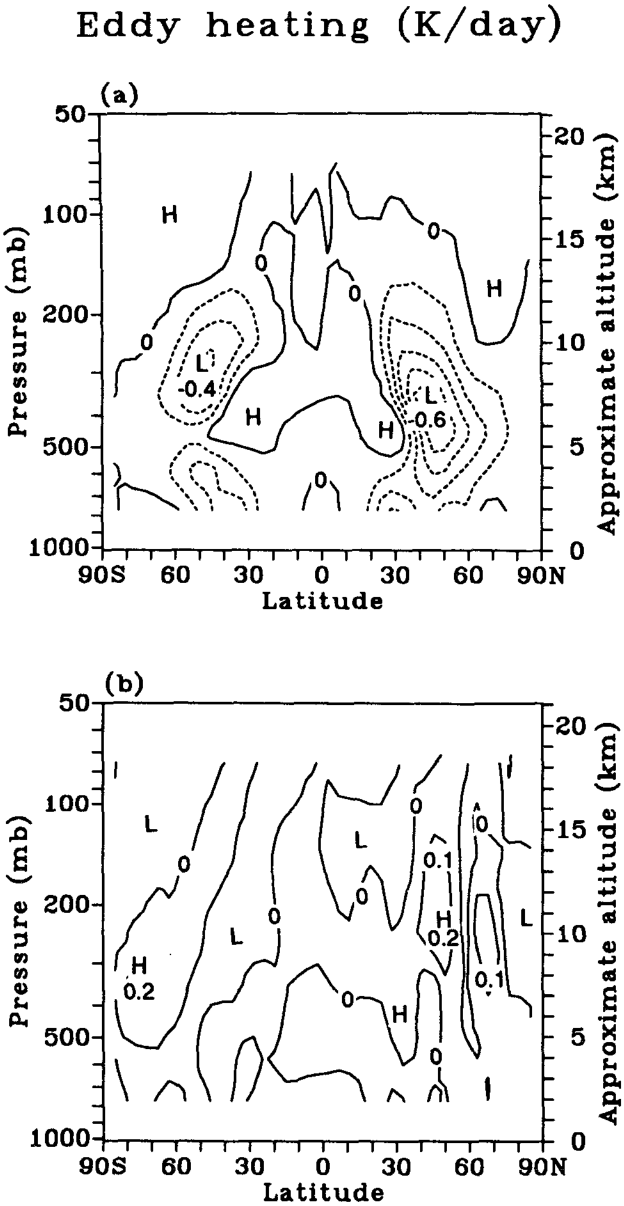

FIG. 2. Eddy heat flux convergence (K day^{-1}) in winter. (a) The contribution from the vertical convergence of the meridional heat flux. (b) The contribution from the vertical convergence of the vertical heat flux. Contour interval is 0.1.

$$e_z = \left(\frac{\partial}{\partial z} + \frac{\kappa - 1}{H}\right)\frac{HT_\omega}{p} = \left(\frac{H}{p}\frac{\partial}{\partial z} + \frac{\kappa}{p}\right)T_\omega, \quad (15b)$$

where $T_v \equiv [\bar{v}^*\bar{T}^*] + [\overline{v'T'}]$ and $T_\omega \equiv [\bar{\omega}^*\bar{T}^*]$. Equations (11), (13), and (15) are used to estimate the eddy heating rates and can be compared to the zonal-mean heating rates q defined by $q \equiv Q(p_s/p)^{-\kappa}$.

The data used for this comparison is Oort's (1983) global atmospheric circulation statistics for 1958–73. This long-time data is supplied in 2.5° intervals in latitude from the South Pole to the North Pole and in 11 vertical levels from 1000 to 50 mb and as 12 monthly means. The isentropic slope from this data for winter, here averaged during December, January, and February, is shown in Fig. 1. The general pattern is the midlatitude maxima at the 600- and 100-mb levels and small slopes in the tropical region and at the 200-mb level. It can be seen that the slopes are much steeper in the troposphere than in the stratosphere.

The eddy heating rates e_y and e_z for the winter time due to the meridional and vertical eddy heat fluxes are shown in Figs. 2a and 2b, respectively. The data quality below 850 mb is poor (A. Oort 1992, personal communication), and their vertical derivatives are even less certain; thus, the results below 850 mb are not considered to be reliable and are not plotted. These figures show the meridional contribution to be larger than the vertical contribution in most of the troposphere. The general pattern of e_y consists of significant midlatitude cooling at the midtroposphere and weak heating in the tropical and polar regions. The contribution from the vertical heat flux is generally found to be negligible in most regions, except for a small peak of heating located at high latitudes. This general pattern is not restricted to winter but is repeated in all seasons. Analyzing the terms in (15) shows that the first terms in each equation are dominating, while the second terms, which are correction factors due to vertical density variations, are of secondary importance.

The sum of e_y and e_z for winter is shown in Fig. 3a. The pattern is not much different from Fig. 2a, again being characterized by midlatitude and midtropospheric cooling and weak heating in the tropical and high-latitude regions. The high-latitude heating comes from the convergence of vertical heat flux. The other three panels of Fig. 3 show the eddy heating in the other three seasons. The location and pattern of cooling and heating do not change. However, the strength of cooling reveals the seasonal change, which shows the maximum in winter and minimum in summer. This seasonal change is clear in the Northern Hemisphere while it is not so obvious in the Southern Hemisphere.

The seasonal variation in the eddy heating rates is better shown in the time–latitude section. Figure 4 shows heating rates averaged between 500-mb and 300-mb levels to obtain a typical seasonal pattern in the midtroposphere. As anticipated in Fig. 3, the Northern Hemispheric cooling distribution shows a strong annual variation. The Southern Hemispheric pattern is more complex, showing large month to month variations. Since the number of observation points in the Southern Hemisphere is smaller than of that of the Northern Hemisphere, the strong month to month variation may be due to a lack of data.

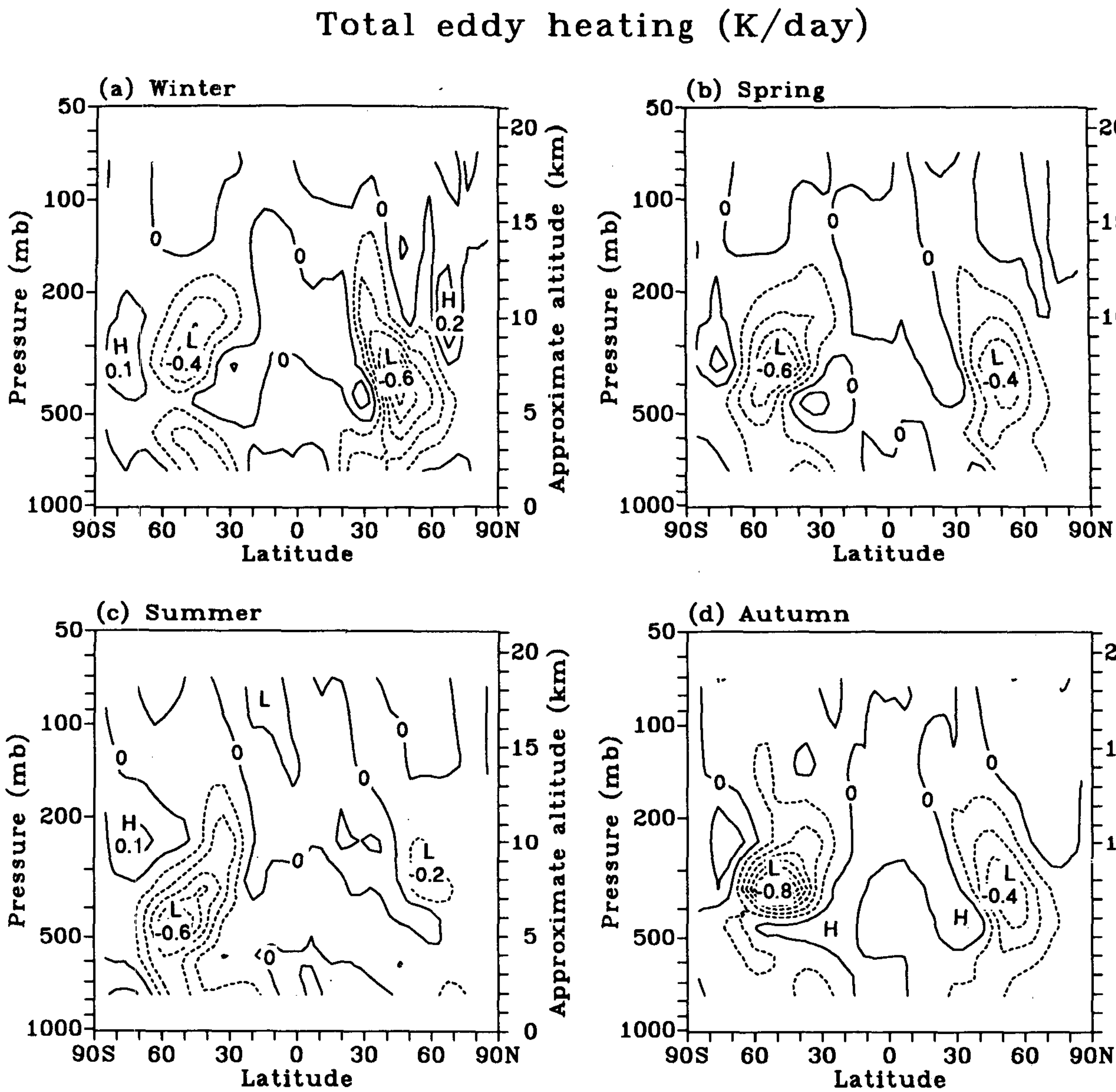

FIG. 3. Eddy heat flux convergence (K day^{-1}) from both meridional and vertical heat flux (a) in winter, (b) in spring, (c) in summer, and (d) in autumn. Contour interval is 0.2.

4. Effects of eddy heating on the meridional circulation

In order to estimate effects of eddy heating, a comparison was done between fields obtained with only zonal-mean heating and zonal-mean heating plus eddy heating. The zonal-mean heating rates used here for each month were obtained from the LLNL two-dimensional radiative chemical transport model. The algorithm for the heating is described in Grant et al. (1987). The heating rates for January are shown in Fig. 5. The main characteristic of these heating rates is the strong tropical heating above 5 km with cooling elsewhere. Comparing Fig. 3a with Fig. 5 shows that at the region where the eddy heating has maximum cooling the strength of the zonal-mean cooling is of the same order of magnitude as the eddy heating. This indicates that cooling due to the eddy effects may significantly contribute to total cooling in the middle-latitude region. The zonal-mean heating rates do not satisfy the global mass balance on certain levels where the global average on isentropic surfaces show net cooling. Since the eddy effect helps to increase global cooling, this term does not help to correct the error appearing in the zonal-mean heating distribution.

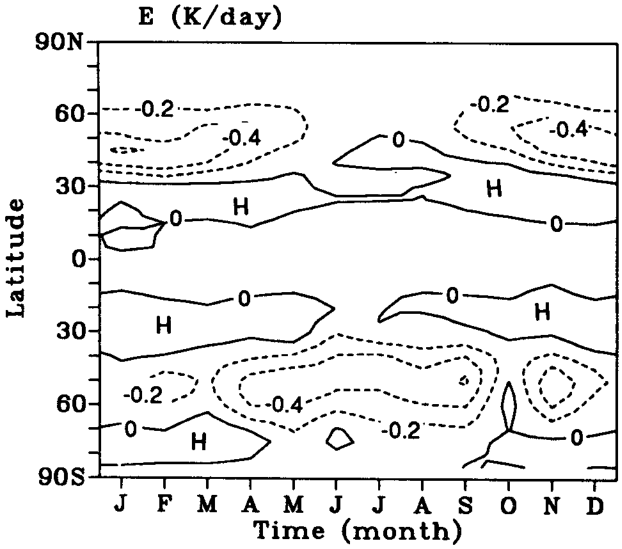

FIG. 4. Time–latitude cross section of the eddy heat flux convergence (K day^{-1}). The values between the 300-mb and 500-mb levels are averaged. Contour interval is 0.2.

The residual mean meridional circulation was obtained through iterations between the thermodynamic energy equation and the continuity equation. After every calculation of the vertical velocity, the global mass balance is obtained by subtracting the globally averaged vertical velocity from every point on the level. This iteration process in the stratosphere needs only five iterations as noted by Solomon et al. (1986),

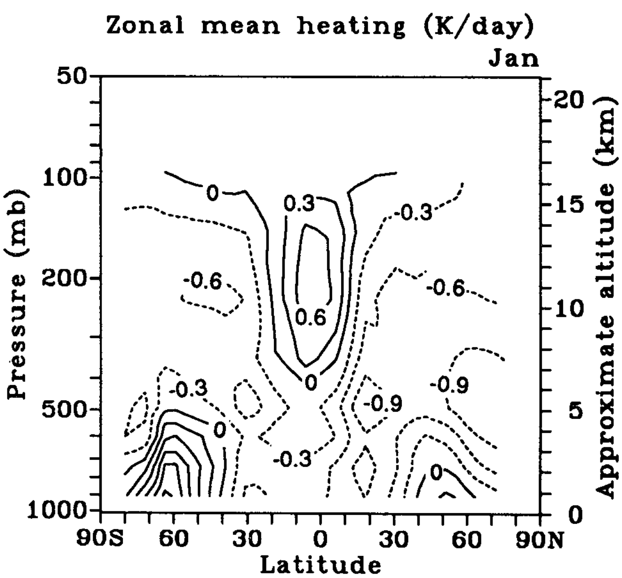

FIG. 5. Zonal-mean diabatic heating rates (K day^{-1}) for January calculated by the LLNL two-dimensional radiative chemical transport model. Contour interval is 0.3.

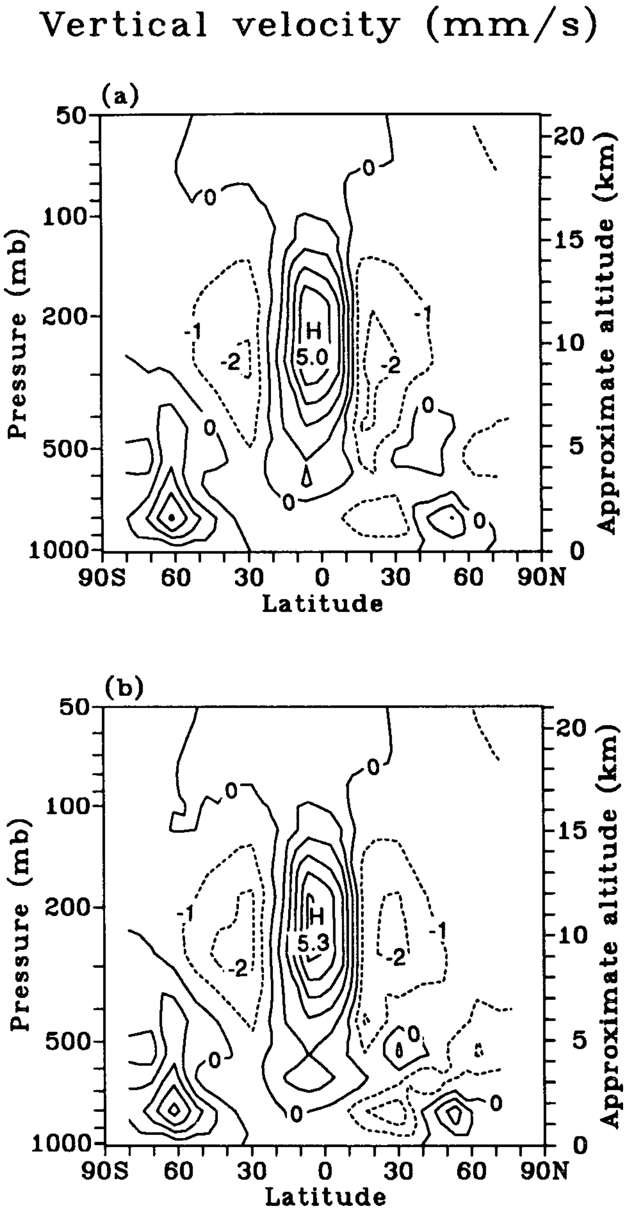

FIG. 6. Vertical velocity (mm s^{-1}) of the residual mean meridional circulation for January calculated based on the zonal-mean heating rates from the LLNL model (a) without eddy heating rates and (b) with eddy heating rates. Contour interval is 1.0.

Hitchman and Leovy (1986), and Gille et al. (1987). In the troposphere, however, more than five iterations are required since the meridional advection of heat is usually not small due to the significant steepness of isentropic slopes. Since the meridional heat advection in the stratosphere is negligible, specifying the zero net radiative heating on each isobaric surface before calculation of the circulation may be allowed, as was done in Rosenfield et al. (1987). However, as noted by Shine (1989), this is not a completely accurate method; in-

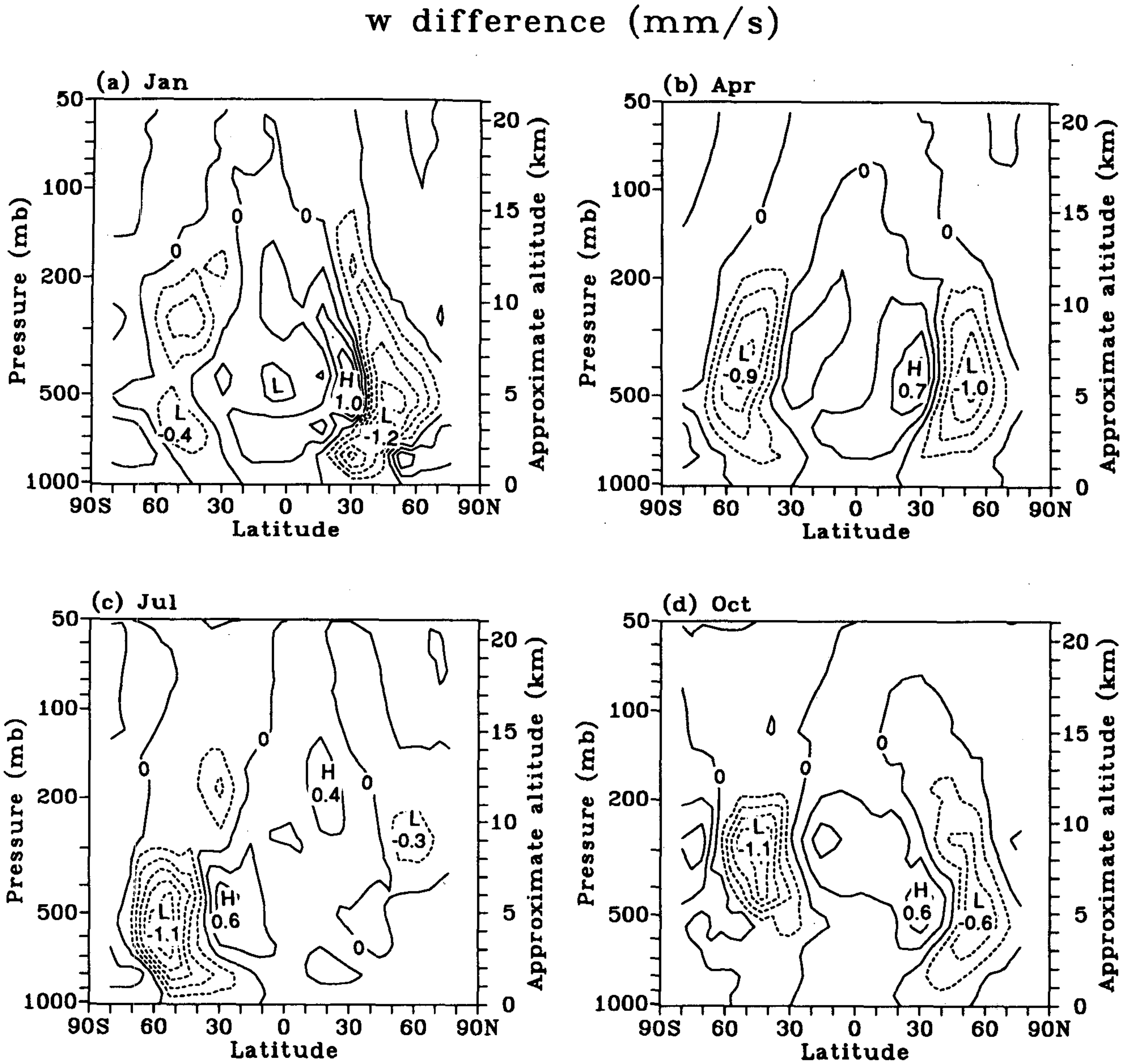

FIG. 7. Differences in the vertical velocity (mm sec^{-1}) of the residual mean meridional circulation for the circulations calculated with and without eddy heating rates in (a) January, (b) April, (c) July, and (d) October. Contour interval is 0.2.

stead, the correct way of adapting the heating rates would be to specify the zero net heating on isentropic surfaces and to interpolate the adjusted heating rates to isobaric surfaces.

The vertical velocity field for January is shown in Fig. 6a. Tropical rising and sinking elsewhere is the dominant pattern with a single-cell circulation pattern in each hemisphere. Since the diabatic heating does not satisfy the mass balance, some regions have velocity fields not corresponding to the heating pattern. For example, there is cooling at the 45-deg and 5-km level in the Northern Hemisphere where rising motion occurs. This is probably a by-product of the calculation procedure that forces a global mass balance at each level.

Another calculation was done for the circulation with the eddy heating added to the zonal-mean heating. Since the month to month fluctuation of the eddy heating rate is very strong in some regions, it was smoothed by the Fourier series. The 12 monthly means of the eddy heating rates were smoothed by retaining the time mean and the annual, semiannual, and terannual components. The eddy heating rates then were interpolated to the model grid points to calculate the circulation. Since the eddy heating is uncertain below the 850-mb level, the value for the lowest model layer (0.75 km) is set to be zero. The result for January is shown in Fig. 6b. The general pattern is similar to Fig. 6a since the eddy heating induces the circulation in the same direction as the zonal-mean circulation. The most significant change is seen at the middle-latitude region in the northern lower troposphere where sinking motion was enhanced. The maximum rising motion at the Tropics is increased even though the eddy heating is negligible for the region. The reason is that the new circulation field calculated with the eddy heating is adjusted by the global mass balance. Figure 3 shows that cooling seems to dominate globally and, thus, enhances the downward motion. However, the artificial adjustment of setting the global net vertical velocity to be zero at every level enhances the rising motion at the equator as a by-product.

The differences between the vertical velocities with the eddy heating and without are shown in Fig. 7 for the four seasons. The rising and sinking pattern is located at the same region annually with varying seasonal magnitude. The magnitudes of the maximum sinking and rising are of the same order of magnitude as that of zonal-mean circulation calculated without the eddy heating in the same region and, thus, may be important. The eddy effects are weakest in summer and strongest in winter in both hemispheres. Figure 8 shows the annual average of the differences in circulation. [Note, before the vector diagram is drawn, the component of the vertical velocity is multiplied by the scale factor, here taken as 950 (=20 012/21.06), to match the vertical stretching of the frame of the figure. Therefore, the length the arrows does not correspond to the real velocity magnitude.] The relative importance of the difference of the circulation by the latitude and altitude is well revealed. Generally, the rising motion is not as important as sinking motion, except near 30° at the 5-km level. This eddy-induced sinking motion is an important contribution to the total circulation and strengthens the single cell in each hemisphere.

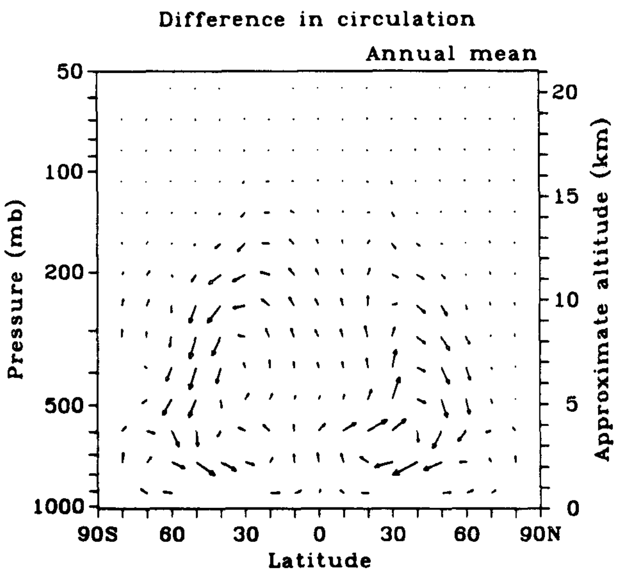

FIG. 8. Annual mean of the difference of the velocity with and without eddy heating rates. The vertical component of the velocity vector was multiplied by the scale factor 950.

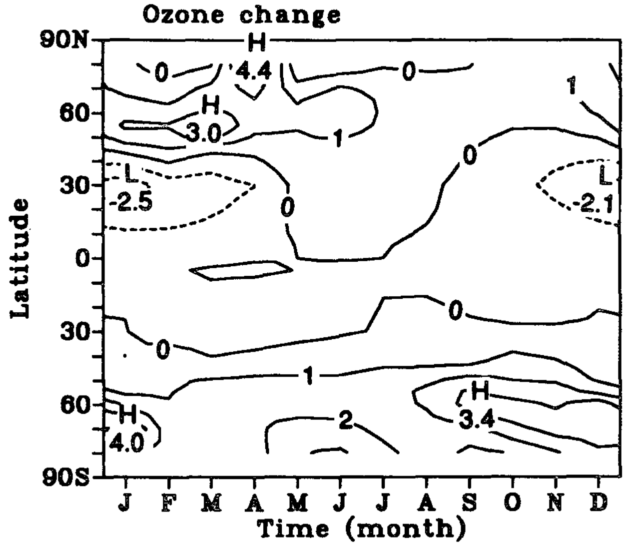

FIG. 9. Time–latitude cross section of the percentage difference in column ozone between model runs with and without eddy heating. Contour interval is 1 percent.

5. Effects of eddy heating on the distribution of column ozone

The LLNL two-dimensional model was used to estimate the change in chemical species distributions caused by the eddy-induced heating. The direction of the circulation difference is persistent throughout a year, and, thus, it may cause significant change of species distribution despite the small magnitude of velocity difference. The LLNL two-dimensional model is described in Johnston et al. (1989), Kinnison et al. (1992), Connell et al. (1992), Patten et al. (1992), and Wuebbles et al. (1993). The model was run until an ambient atmosphere was established with and without eddy heating rates. We compared mixing ratios of nitrous oxide, methane, and ozone between the two ambient atmospheres.

The change in mixing ratio of a chemical species depends on the advection velocity and the gradient of mixing ratio. Since the mixing ratio gradient of nitrous oxide and methane in the troposphere and lower stratosphere are small, the change in mixing ratio is insignificant, with the biggest change of a few percent (not shown). The gradient of ozone mixing ratio, however, is fairly large near the tropopause region and in the troposphere. The time–latitude representation of the percentage difference in column ozone between the model runs with and without eddy heating is shown in Fig. 9. The general pattern shows a decrease in lower latitudes and an increase in higher latitudes. In the Northern Hemisphere the maximum decrease occurs at 30° latitude in winter in which rising motion due to the eddy effect is strongest (Fig. 7a). However, the maximum increase occurs in high latitudes a few months later in spring, despite the maximum sinking during winter. This implies that there exists a difference in timescale of transport between the rising branch and the sinking branch. The pattern in the Southern Hemisphere is slightly different from that in the Northern Hemisphere. A decrease in the column ozone at 30° latitude does not appear, and, also, the period of ozone increase in the middle latitudes lasts until summer.

The addition of this eddy heat flux term provides a total ozone distribution in better agreement with observations. [For a complete description of total ozone distributions of the LLNL and other models see the report of the National Aeronautics and Space Administration Model and Measurement Workshop, 1992; for reference, see Wuebbles (1993).] As described, the addition of the eddy term tends to decrease ozone in the lower latitudes and increase ozone in the middle to upper latitudes. This mid- to high-latitude increase in ozone is especially evident during the winter months of each hemisphere. These regions of increases correspond to the regions where the LLNL 2D model has tended to underestimate the column ozone, especially in the Southern Hemisphere winter. The addition of the eddy heating gave a model representation of the ozone maximum between 60° and 70°S that more closely resembles observations.

6. Summary and discussion

The eddy heat flux convergence, found as a forcing term in the thermodynamic energy equation, was estimated from the long-time climatological data compiled by Oort. This term has been neglected in previous calculations of the residual mean meridional circulation. The estimation shows that the eddy heating rates may be important in the troposphere, especially in the midlatitude midtroposphere. The general pattern is in the same sense as the zonal-mean heating rates rather than acting to cancel the mean heating. It has significant cooling in the middle latitude up to about 0.6 K day^{-1} and weaker heating in the low latitude and, thus, strengthens the cooling and heating of the zonal-mean heating rates. The contribution for the eddy heating from the meridional eddy heat flux is dominant over the vertical heat flux.

The meridional circulation with and without eddy heating effects was calculated by the iteration between the thermodynamic energy equation and the continuity equation. The effects of eddy heating on the circulation is an enhancement of the zonal-mean circulation. It strengthens the sinking motion up to about 1 mm s^{-1} and increases the tropical rising motion. Thus, the circulation induced by the eddy effects is a single-cell circulation in each hemisphere as in the zonal-mean circulation. A second cell appears in the high-latitude region; however, the magnitude of the velocity related to the circulation is very small.

The LLNL two-dimensional model was used to estimate the effect of eddy heating rates on the distribution of various chemical species. Very little change is seen in the distribution of species whose spatial gradient is small in the troposphere and lower stratosphere. However, in the case of the column ozone change eddy effects contribute to a decrease in lower latitude and a larger increase in the higher-latitude column ozone.

The eddy heating rates and their effects on the circulation and tracer distribution are a significant feature within the troposphere. However, their importance on the calculation of the circulation depends on the accuracy of the consideration and treatment of zonal-mean heating rates. With current-generation chemical transport models of the global atmosphere now basing calculations on derived heating/cooling rates, it is important that the eddy heating rates be properly considered.

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