EXPERIMENTS WITH A STRATOSPHERIC GENERAL CIRCULATION MODEL

II. LARGE-SCALE DIFFUSION OF TRACERS IN THE STRATOSPHERE

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ABSTRACT

The 18-level primitive equation, general circulation model described in Part I was used to study the diffusion of two idealized tracers in the stratosphere. One tracer was designed to simulate broadly the behavior of the radioactive tungsten which escaped into the stratosphere following nuclear tests in the Tropics, the other was taken as a photochemical ozone distribution. Both the meridional circulation and the large-scale eddies were found to be important in the diffusion of the tracers, and for quasi-steady state conditions they formed a highly interrelated system in which their actions were mutually canceling. The large-scale eddies were of primary importance for the polewards transport of the tracers in middle and high latitudes, but the supply of tracer for these eddies was principally maintained from the higher levels by the downward branches of the meridional circulation. Two meridional cells were found to occur in the stratosphere, a tropical direct cell and a higher latitude indirect cell, and these provided a natural explanation for many of the observed features of the tracer distributions in the actual atmosphere. The only major tropospheric-stratospheric exchange took place in the subtropics through the tropopause gap, the vertical eddies and the meridional circulation being of comparable magnitude for this exchange.

The synoptic situation in the atmosphere was found to be of fundamental importance for the large-scale diffusion of the tracers in middle latitudes, and the downgradient transport of tracers in the lower stratosphere was primarily accomplished by the upper level troughs of the planetary scale wave system.

Although the model used in this investigation was based on radiative conditions corresponding to annual mean insolation it appeared to be representative of winter conditions, and was in agreement with many observational features.

Schematic diagrams illustrating the principal features of the large-scale diffusion of the two tracers are given in figures 12 and 24.

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1. INTRODUCTION

The ability of the atmosphere to transport certain quantities has long been known, from such rare phenomenon as the Krakatoa explosion to the commonplace yearly variation of the ozone amount. The manner in which such quantities are distributed over the globe, and in particular transported across lines of latitude, has always been a question of considerable interest, which perhaps culminated in recent years in the study of the radioactive debris deposited in the atmosphere from the testing of nuclear weapons. The movement of ozone and radioactive debris is primarily of interest in the stratosphere, as once they enter the troposphere they have a lifetime only of the order of 1 mo. because of washout and deposition. On the other hand, their lifetimes in the stratosphere range from several months to years, so that these substances become a quasi-conservative property of the flow, and this has led to the concept of using them as tracers to deduce the atmospheric motions of this region.

Because the basic flow pattern in the atmosphere is zonal, the major problem with tracers is to account for their meridional distributions and the variation of their concentrations with height. Of these the former has been investigated in most detail and two principal methods of explaining this transport have been proposed. One of the early suggestions by Brewer [3] and Dobson [8] was that the meridional circulation dominated the transport. However with the gradual accumulation of knowledge concerning the role of eddies in the atmosphere, which transport indirectly by means of correlations between the flow and tracer fields, it is now known that in middle and high latitudes eddies primarily control the transfer of ozone and radioactive tracers in the lower stratosphere (see for example Newell [29]). The meridional cell is still thought to predominate in the Tropics, but the relative magnitudes of the two mechanisms are rather poorly known because of the scarcity of data at stratospheric levels, and much remains to be done before we can claim to understand the transport processes in the stratosphere and above.

Most of the previous numerical work on the transport of tracers in the stratosphere has been based on two-dimensional models, in which it is assumed that the transport is governed by eddy diffusion alone. Davidson et al. [6], Machta [24], and Reed and German [34] have all used this type of model to study the diffusion of radioactive debris in the stratosphere. Prabhakara [32] has attempted to explain the latitudinal distribution of ozone with a hybrid model, which consisted of the usual form of two-dimensional eddy diffusion model but also included the effects of mean meridional transports by the addition of suitable velocities. These models are, of course, highly parameterized, since it is necessary to specify somewhat arbitrary eddy diffusion coefficients for the horizontal and vertical transports. The only previous three-dimensional study appears to be that of Byron-Scott [4], who investigated the transport of ozone by using a model with equations scaled to the stratosphere, the flow in the stratosphere being forced by the supply of energy at the lower boundary of the model.

An investigation of these transport problems using a three-dimensional general circulation model therefore seems to be particularly appropriate. With such a model it is not necessary to resort to parameterization in order to obtain transport by eddies, since the model generates its own eddies in the same manner as the atmosphere. In addition the model also develops its own meridional circulation, and thus both the major transport mechanisms in the atmosphere are intrinsic properties of the model. Apart from just reproducing the diffusion of the tracers with the model, a feature of great importance is that one can at least attempt to answer some questions concerning which mechanisms govern the processes resulting in the calculated distributions, something which was not possible with previous models. Some of the possible questions of interest are: What are the relative roles of the meridional circulation and the eddies in transporting a tracer, both as a function of altitude and latitude during the life cycle of that tracer, and how are these two mechanisms related? Do the potential temperature surfaces constrain the movement of the tracer, and what is the dynamic and radiative situation which permits the tracers to cross these surfaces? Why do tracer distributions, in general, have a marked discontinuity in their vertical profiles in the vicinity of the tropopause, and why are concentration gradients in the equatorial stratosphere crowded against the tropopause? What mechanism governs the transfer of tracers from the stratosphere to the troposphere, i.e., large-scale eddies, local diffusion, direct transfer through the tropopause gap?

The present work was undertaken in the hope that some of these questions might be answered by trying to simulate the transport of tracers initially distributed in the stratosphere. Two tracers were investigated, the first was designed to represent a nuclear bomb test in which radioactivity was primarily deposited in the equatorial stratosphere; while with the second it was hoped to get some preliminary insight into the transport of ozone. In the case of ozone there is an additional complicating factor to be considered since ozone is created photochemically in the atmosphere. Photochemical theory actually predicts maximum total ozone amounts in the Tropics, in disagreement with observation (see Craig [5]), and it is generally agreed that large-scale advection is responsible for the difference between theory and observation. In order to understand how the large-scale motions accomplish this, two basic approaches are possible:

1) An experiment can be started from a photochemical equilibrium ozone distribution, and, by excluding any subsequent photochemistry and treating the ozone as an inert tracer, the subsequent modification of the initial distribution can be followed. This approach permits the dynamical effects to be isolated as regards their role in arriving at the observed ozone distribution, and it is particularly desirable when the quantitative aspects of the photochemistry are ambiguous (see Hunt [13]).

2) The same initial conditions as in experiment 1) could be used, but with the photochemistry incorporated, so that a joint photochemical-dynamical integration would be performed. Ideally this should result in an ozone distribution in agreement with observation, and the combined actions of the photochemistry and dynamics in maintaining this distribution could be analyzed.

As a first step experiment 1) was undertaken and will be discussed in this paper; experiment 2) is currently underway.

2. INITIAL CONDITIONS AND TIME INTEGRATION

Only those aspects related to the tracers will be given here since a description of the dynamical part of the model has been given in Part I, Manabe and Hunt [26]. The number of tracers which could be incorporated in this experiment was limited to two by the core size of the computer. Ideally one would have preferred to have included a larger number of tracers so that a greater variety of initial conditions and types of tracers could have been investigated, especially since the addition of tracers to the model produces only a very small increase in computation time.

Compared to any situation likely to exist in the atmosphere the two tracers studied with the model were rather idealized. This was done deliberately in order that the behavior of generalized, rather than specific, tracers could be investigated, as the information content of such an investigation was considered to be higher. The initial conditions of the radioactive tracer, hereafter referred to as

TABLE 1.—*Heights and pressures of the model layers*

Level	Height (km.)	Pressure (mb.)
1	37.5	4.0
2	29.5	12.9
3	25.6	23.4
4	22.8	36.1
5	20.5	51.2
6	18.5	69.4
7	16.8	91.1
8	15.2	117.1
9	13.7	148.2
10	12.3	185.5
11	10.9	230.1
12	9.55	283.6
13	8.20	347.5
14	6.75	424.1
15	5.35	515.8
16	3.9	625.6
17	2.4	757.0
18	0.85	914.3

R1, consisted of defining a distribution which was symmetric with regard to level and which was centered on level 5, and located as a uniform zonal ring around the Equator; see figure 4a (note this figure shows concentration versus pressure not level). As indicated in the figure the initial distribution was also symmetric about the Equator, even though the wall at the Equator in the model meant that the distribution terminated at the wall. This symmetry about the Equator was designed to remove any equatorial transport of tracer at the start of the experiment. The initial mixing ratios of R1 were given values at the Equator of 2 at level 5, 1 at levels 4 and 6, and ½ at levels 3 and 7, all other levels being set equal to zero. The pressures and corresponding standard heights of these levels are given in table 1. The latitudinal extent of the R1 zonal ring was approximately 10°, the mixing ratios reducing from 2 to 1 to ½ to zero at consecutive grid points at level 5 and over the appropriate range for the other levels. No particular significance should be attached to the values assigned to the initial mixing ratios, as they were primarily chosen to prevent excessive truncation error at the beginning of the integration. Similarly the units assigned to R1 of gm./gm., which were the simplest possible, have no significance and μ gm./gm. or gm./kg. could equally well have been used. These initial conditions bear no relation, of course, with the manner in which radioactive debris is placed in the stratosphere by a nuclear explosion, since this corresponds to a point source. The choice of a uniform zonal ring does, however, have the advantage that any subsequent perturbations observed in the tracer pattern can be directly attributed to atmospheric motions, rather than asymmetries in the initial concentrations, and this is useful in interpreting the results.

The second (ozone) tracer, referred to as R2, was defined initially as a photochemical ozone concentration, and was based on the oxygen-hydrogen atmosphere model discussed previously by Hunt [14]. The initial R2 values were calculated at 2-km. intervals in a separate computation and were based on the 1962 U.S. standard atmosphere. Calculations were made for latitudinal intervals of 5°, the solar zenith angle at each latitude being that obtained by averaging over the daytime in the manner described by Manabe and Möller [27]. These values were then interpolated linearly in order to define initial R2 values at every grid point for the top eight levels of the model. At lower levels, below approximately 14 km., R2 was taken as zero initially. Since no photochemistry was included in the present model it was obvious that the R2 distribution obtained at the end of the experiment could not be representative of actual atmospheric ozone concentrations.

The time variation of the tracers in the model was governed by the following equation:

$$\frac{\partial}{\partial t}(p_* r) = -m^2\left[\frac{\partial}{\partial X}\left(\frac{p_* U r}{m}\right)+\frac{\partial}{\partial Y}\left(\frac{p_* V r}{m}\right)\right] - p_*\frac{\partial}{\partial Q}(\dot{Q} r) + {}_H F_r - p_* C$$

where r is the tracer mixing ratio, ${}_H F_r$ is the flux divergence associated with the subgrid scale horizontal diffusion, C is the "precipitation" of the mixing ratio, and p_* is the surface pressure. The stereographic map factor is m, U and V are the velocity components corresponding to the X and Y coordinate directions respectively on the stereographic projection, and $\dot{Q}$ is the vertical velocity in the Q coordinate system (see Part I). This equation was an integral part of the complete general circulation model, and was integrated simultaneously with the basic equations defining the dynamic properties of the model. The dynamic and diffusion studies therefore proceeded concurrently, and no restraints whatsoever were applied to the diffusion of the tracers by the model dynamics, and the presence of the tracers in no way influenced the dynamics.

The eddy and mean meridional transports of the tracers which occurred were simply generated by the large-scale diffusion processes associated with this equation. The former are obtained by the conventional breakdown of the transport into two terms involving a mean, and a departure from that mean. The eddy component obtained in this way is the basic Reynolds' stress term, which in the two-dimensional eddy diffusion models is parameterized as an eddy diffusion coefficient and a mixing ratio gradient. Recourse to this procedure is obviously unnecessary in the general circulation model. The eddy terms referred to here are not quite the same as those known as transient eddies in observational studies of the atmosphere. The latter are obtained for an individual station for a specified time period by evaluating them as the difference, $\overline{v'a'}^t$, between the time-meaned total flux, $\overline{va}^t$, and the flux derived from the product of the time-meaned individual terms, $\bar{v}^t\bar{a}^t$, where v is an appropriate speed and a is any term of interest. If sufficient stations exist in a given zonal belt the results can also be averaged to provide a representative zonal mean, $\overline{\overline{v'a'}^t}^\lambda$. In the case of the model, for a given instant of time and for an appropriate latitudinal range, it is more convenient to work in terms of a zonal mean, $\bar{v}^\lambda\bar{a}^\lambda$, and a departure from that mean, $\overline{v\,a'}^{\lambda}$. The results may then be time averaged over a suitable period to produce a representative zonal mean eddy flux $\overline{\overline{v'a'}^{\lambda}}^{t}$. The fluxes obtained in this way will be referred to simply as large-scale eddies.

In view of our ignorance concerning the particle sizes involved and the magnitude of vertical diffusion, no allowance was made for either subgrid scale vertical diffusion or gravitational settling of the tracers. Precipitation was arbitrarily imposed by removing immediately any tracer which reached the 16th level (626 mb.) or below, the precipitation being accumulated at the surface. This approach implied that washout by rain or destruction of the tracers was zero above this level and 100 percent efficient below. The influence of the latitudinal variation of rainfall intensity was implicitly ignored in this way, even though it is known that radioactive deposition is affected by rainfall (Junge [15], p. 261). In the case of ozone, precipitation in the atmosphere would appear to be of minor importance compared with destruction at the surface, which is the principal sink for ozone (Junge [15], p. 56).

The basic hypothesis underlying the above equation, and thus the whole experiment, was that the mixing ratio acts as a conservative quantity at all times and places. As far as the actual atmosphere is concerned this assumption appears to be valid for the stratosphere, but would be expected to fail in the troposphere at points where losses were high. The numerical scheme conserves the mass integral of the mixing ratio, but not the mixing ratio of an individual air parcel in the Lagrangian sense due to truncation error, which accounts for some final R2 values in level 1 being greater than any initial values. In addition this truncation error also affected some individual points, mainly at lower levels, where at times negative mixing ratios were obtained. These negative values were reset to zero and the surrounding points had their concentrations reduced to counteract the negative value. In cases where the concentrations of those points were too low to cancel entirely the negative value this resulted in a small fictitious source of mixing ratio. Ordinarily the adjustment mechanism gave only very small contributions, and as the tracer patterns developed the adjustment became less frequent because the distributions were smoother.

The mixing ratios were initiated in the model 84.5 days after the start, when it was thought that the model had settled down following its adjustment to the initial conditions. Subsequently it appeared that the stratosphere had not yet reached a quasi-equilibrium, dynamic state, and, as discussed in Part I, the top layer in particular underwent some dramatic changes. This did not seem to result in any marked changes as far as the tracers were concerned, presumably because this layer is not important for R1, while in the case of R2 the latitudinal concentration gradient is quite small. The tracers were diffused by the model for a total of 180 days when the experiment was terminated, even though the tracer distributions were still changing quite rapidly.

The remainder of this paper will be concerned with discussing the transport of the tracers. Since the initial distributions of R1 and R2 were radically different they were influenced in different ways by the atmosphere; hence their life histories will be presented separately for convenience. For the purposes of this discussion zero time will normally be taken to be the time at which the tracers were initiated.

3. DIFFUSION OF THE RADIOACTIVE TRACER, R1

3.1. GENERAL DESCRIPTION

In this section a brief discussion will be given of the diffusion of R1 in order to provide an overall pictorial background before giving the numerical details of the transport mechanisms. This discussion is largely based on a series of hemispheric maps made for the various levels during the course of the experiment. The maps gave instantaneous distributions of the tracer concentrations and were made at a frequency of 4 days, which was considered adequate to resolve the features of the diffusion. Although these maps are extremely useful in visualizing the movement of the tracer, because of space limitations, it is only possible to present a very limited sample. The sample selected was designed to reveal some of the most representative features, and consists of a time sequence of 5 different days for level 5 in figure 1, and a height sequence for five different levels for one time in figure 2.

However before discussing these figures a gross indication of the change in the vertical distribution of R1 resulting from the transport processes is given in figure 3a, in order that the adjustment of the total amount of tracer in a horizontal level can be seen in perspective. The figure clearly indicates that the net result of the transport has been to increase the concentrations in level 4 and above at the expense of level 6 and particularly 5, while level 7 and below have also increased to a lesser degree. It should be pointed out, though, that figure 3a is slightly misleading because of the way it is plotted. In order to see the exact changes which takes place in the R1 distribution it is necessary to plot the hemispheric content per level weighted with the mass of its corresponding layer. If this is done the curve for final conditions, while having the same shape and height of the maximum, shows an approximate doubling of the values for the lower layers, 7–10. Figure 3a also reveals that essentially all of R1 remained in the middle of the stratosphere where it was initially placed. On the other hand figures 1 and 2, while indicating the horizontal spread of the tracer, give the impression that overall the total concentration of a level would be expected to have decreased quite markedly from its initial value. In figure 3b the latitudinal variation of the vertically integrated R1 amount is given for various times. The change between the initial and final conditions is quite remarkable, and the final distribution is more uniform than one might have expected.

Regarding the horizontal variation of R1, one might have expected a fairly uniform zonal distribution to have been maintained while a general poleward progression was observed. This is not at all the picture which emerged

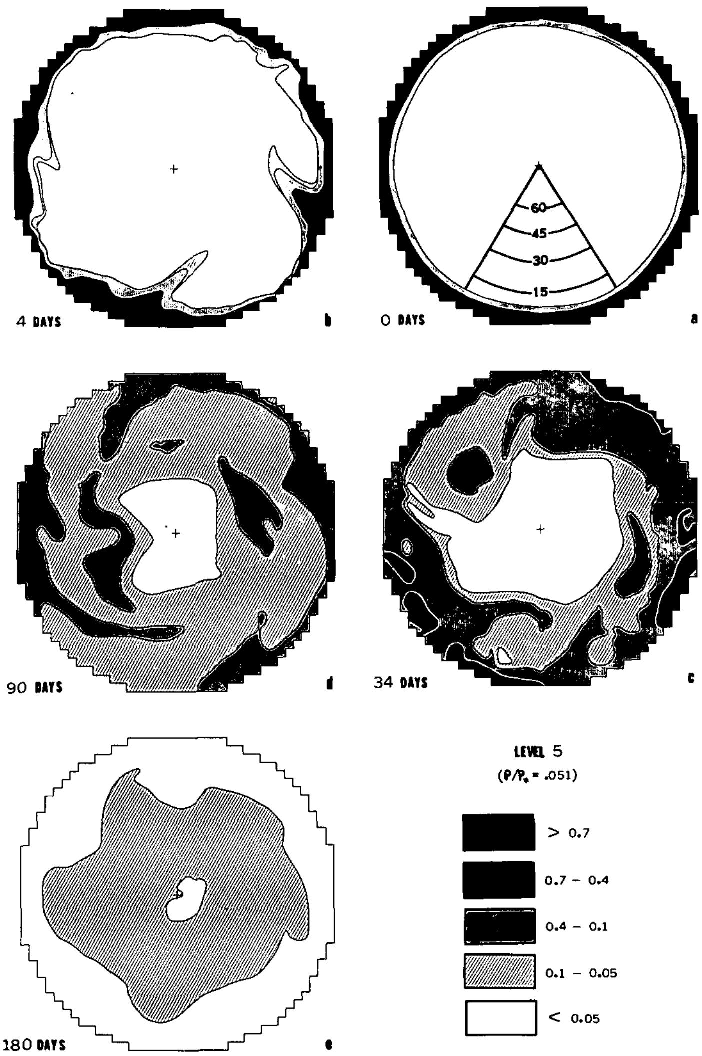

FIGURE 1.—Hemispheric variation of R1 at level 5 for various instants of time. (Units: gm./gm.)

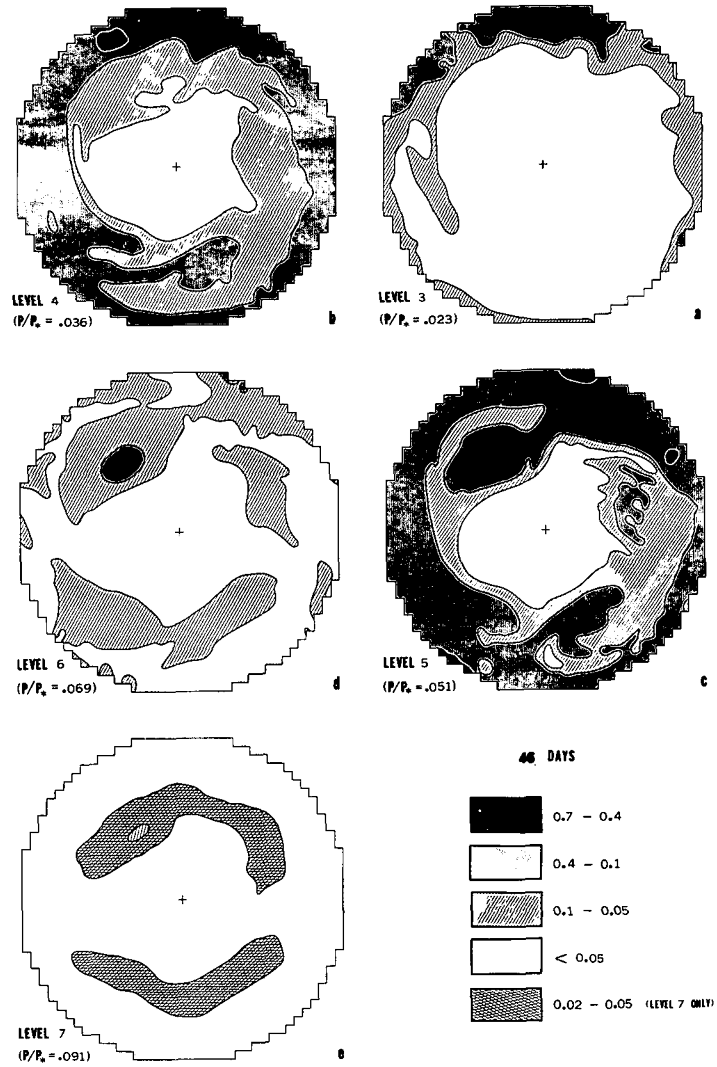

FIGURE 2.—Hemispheric variation of R1 for principal levels 46 days after initiation. (Units: gm./gm.)

in the early stages, as figures 1 and 2 reveal that quite marked asymmetries were produced in the zonal tracer distribution in spite of the fact that the flow is predominantly zonal. In any given level perturbations of the initially symmetric distribution appeared to develop at four locations, although only two were active at any given time, and this may be related to the fact that wave number 4 predominates in the vicinity of level 5 (see fig. 4D1 of Smagorinsky et al. [35]). The perturbations that grew from these points consisted in general of a long neck or promontory of higher concentration relative to the surroundings, which in some cases extended several thousand kilometers and had a characteristic tilt which was primarily in the zonal direction. Typical examples are shown in figures 1 and 2.

Marked development of perturbations can be seen at two places in figure 1 within 4 days of initiating the tracer into the model. As might be expected these perturbations were most prominent in the early stages when the concentration gradient of the tracer was highest. The growth of perturbations seemed to go through a cycle of about 2 or 3 weeks with a history roughly as follows. A promontory of tracer grew as a long narrow neck of high concentration relative to its surroundings until eventually the neck snapped, leaving an "island" of high concentration. These islands disappeared either by diffusion or attachment again to the main tracer belt at some point other than where they were initiated. A new promontory would then start to develop from the remains of the old one. The existence of these promontories and islands is not inconsistent with observations, as secondary maxima have been observed in the concentrations of radioactive tungsten in the stratosphere (Stebbins [36]). These measurements were made principally by U-2 flights along a line of constant longitude, and the secondary maxima reported can be interpreted as resulting from the flight path intersecting a promontory. It should be recalled that the model had a uniform zonal distribution of R1 at the start and also had no land-sea contrast or orographic features, and in spite of this very marked asymmetries developed in the tracer pattern. This would therefore suggest that asymmetries in observed tracer distributions may arise from hydrodynamic causes besides those due to standing eddies associated with topographical features of the lower bound-

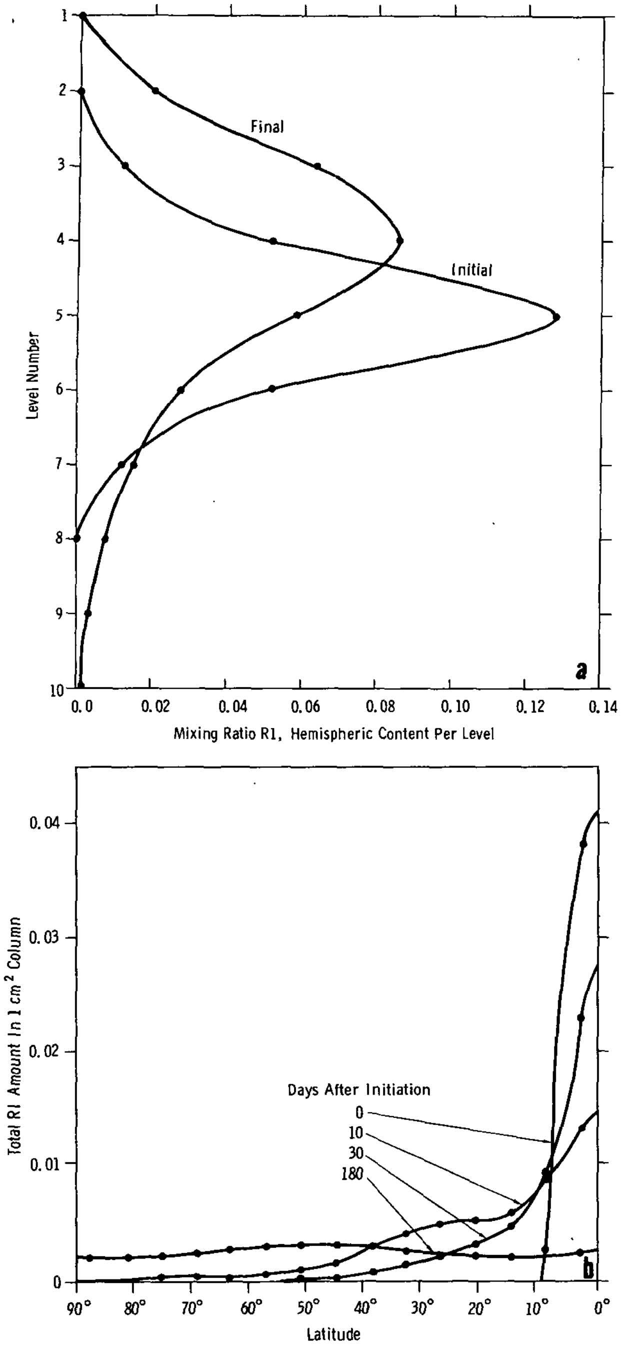

FIGURE 3.—In the upper part of the figure the hemispheric integral of the R1 amount in different levels is illustrated for initial and final conditions. In the lower part the latitudinal variation of the vertically integrated R1 amount in a column of the atmosphere is shown for various times. (Units: gm./(gm. cm.²).)

ary of the atmosphere. The cause of these asymmetries is presumably the waves produced in the lower stratosphere by the forcing of the baroclinic instability in the troposphere.

Some details of the behavior of the individual layers will be given to complete this section. Only the levels in which R1 was put initially were studied in detail. Levels 3 and 4 exhibited quite different modes of diffusion at the start, since the former rapidly redistributed its tracer resulting in maximum concentrations appearing at the four "corners," while the latter broke down its concentration more gradually and at all times had a continuous equatorial belt of maximum concentration which was maintained by transport of R1 from level 5. For about the first 85 days there was a slow poleward progression of this equatorial belt in level 4, which at its maximum extended from the Equator to approximately 30°. However as the concentration in level 5 became more reduced this progression ceased and the belt gradually contracted. At the end of the experiment even this equatorial belt was beginning to break down, as level 5 had insufficient tracer to maintain it, and it became reduced to a series of spiral bands essentially zonally oriented and emanating from the Equator. On the other hand level 3 gradually accumulated more tracer and lost its early, patchy appearance, and after about 100 days also had a continuous equatorial belt of maximum concentration, which was maintained from level 4 by the meridional circulation. By the time the experiment was terminated level 3 had a considerably more uniform tracer distribution than level 4. There thus appears to have been a steady upward progression of the equatorial maximum, which in the case of the actual atmosphere should lead to longer residence times permitting a greater loss of radioactivity by natural decay.

As might be expected of the level with the maximum initial concentration gradient, level 5 showed the maximum variability and developed the biggest perturbations in the early stages. After about 60 days it became rather quiescent and gradually broke up more and more. The concentration also decreased continuously and there was a general movement of the maximum of the concentration from the Equator to middle latitudes, the poleward development being considerably faster than in level 4. At the end of the experiment, as shown in figure 1, level 5 had a remarkably uniform mixing ratio of between 0.1 and 0.05 except at the Equator. This should be compared with its initial maximum value of 2 at the Equator where a minimum now exists.

Both levels 6 and 7 had their initial mixing ratios very rapidly reduced, the former going from a maximum of 1.0 to isolated values of around 0.1 within 30 days, while the latter was reduced from 0.5 to 0.1 within 10 days. The concentration in level 6 then tended to stabilize with a maximum value of 0.05—0.1. After about 65 days the concentration pattern consisted of an almost complete annulus centered at about 40° lat.; figure 2 shows the early stages of the buildup of this annulus. The annulus remained fairly well fixed at this latitude and rotated anticlockwise with a period of about 3 weeks. The growth of concentration in level 6 resulted in a similar growth in level 7 at middle latitudes, individual features of the concentration pattern being recognizable in both levels. However, unlike 6 the tracer in level 7 gradually extended to the Poles, so that by the end of the experiment a fairly uniform polar cap extended down to about 30° lat. and clearly exhibited wave number 6. At the end both levels 6 and 7 had minimum concentrations at the Equator which were essentially zero.

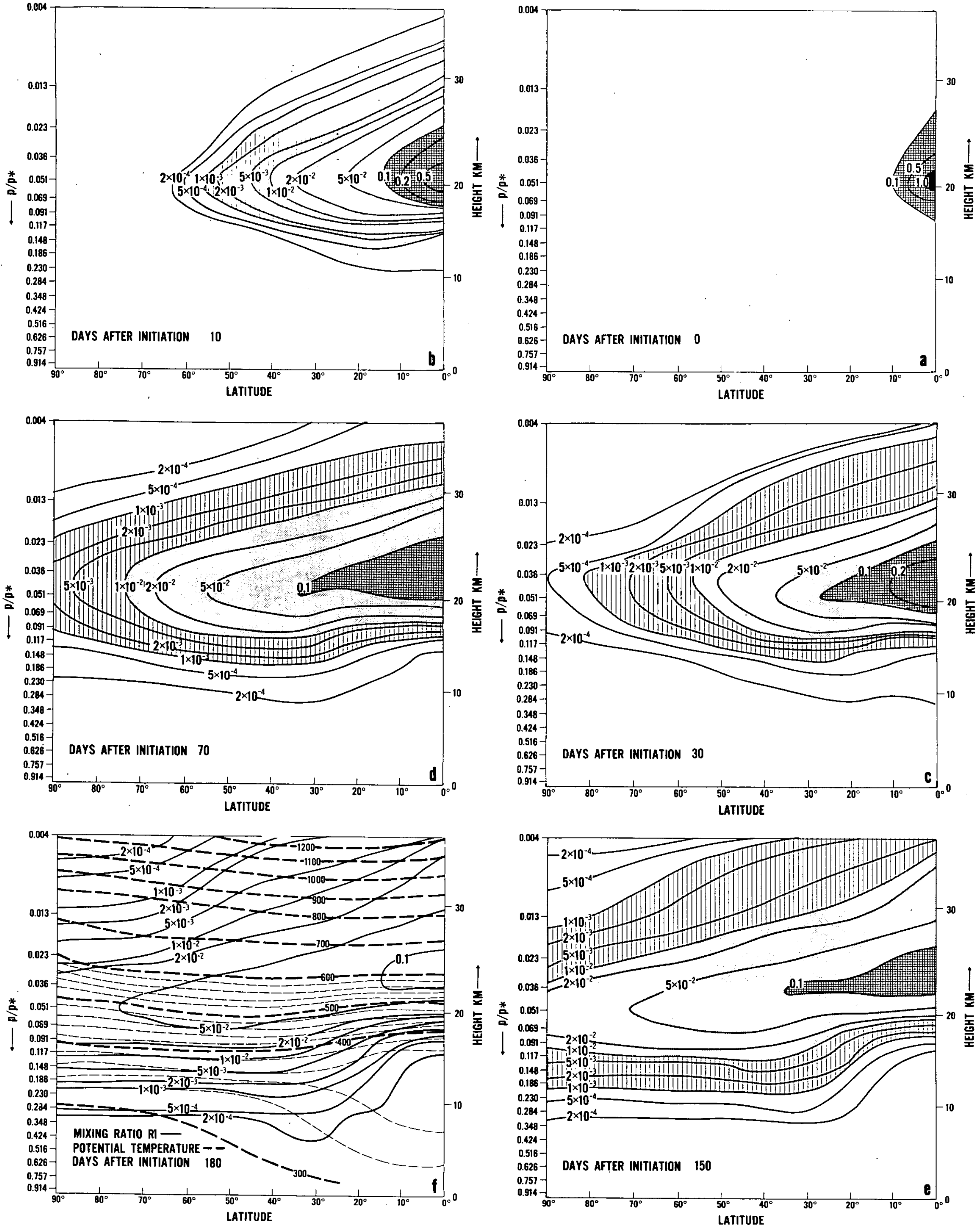

FIGURE 4.—Latitude-height distributions of the zonal mean R1 concentrations for various times. (Units: gm./gm.)

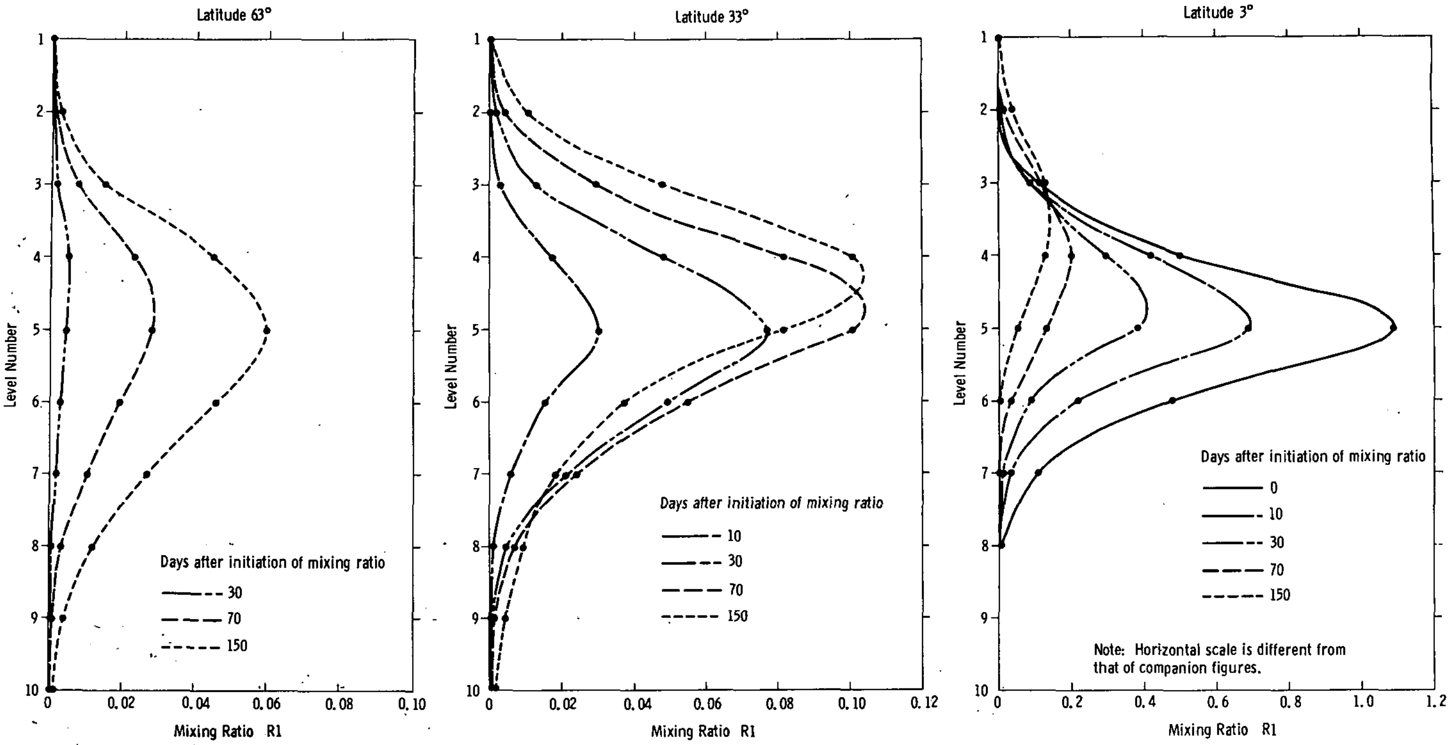

FIGURE 5.—Vertical R1 distributions for various times at different latitudes. (Units: gm./gm.)

3.2. LATITUDE-HEIGHT DISTRIBUTION

The description of the transport mechanisms in the succeeding sections will be greatly facilitated by presenting latitude-height distributions of R1 first. In figure 4 such distributions are given for zonally averaged conditions for a "doubling" time interval after the R1 initiation, these distributions being instantaneous, not time averaged. It is obvious that compared with figures 1 and 2, a very large amount of detail in the tracer distribution has been removed by this zonal averaging process. It should also be noted that the isopleths in figure 4 vary by four orders of magnitude, and the tracer therefore does not in reality diffuse as fast as a first glance might indicate.

Probably the most dominant feature of figure 4 is that at all times the maximum concentration remained at the Equator, but showed a gradual increase in height. The polewards diffusion took place essentially horizontally, although there was a very small downwards slope superimposed. The lower stratosphere filled up first, with the highest concentrations at the lowest levels occurring at about 30° lat. in the vicinity of the tropopause gap; as can be seen in figure 4f the maximum concentration in the lower stratosphere progressively moved polewards for increasing heights. A very pronounced feature in figure 4 is that after about 70 days the tracer concentrations in the tropical troposphere were steadily reduced, with the result that the R1 isopleths rose very fast on the equatorial side of the subtropics, and tended to be crowded in the vicinity of the tropical tropopause. The concentrations were extremely small at all times throughout the troposphere and also in the middle polar stratosphere, levels 1–3, but elsewhere varied quite uniformly from the maximum at the Equator.

A somewhat more detailed indication of the consequence of the R1 diffusion is given by the latitudinal cross sections in figure 5. These show more clearly the fairly rapid reduction in the concentrations in the Tropics, particularly at level 5, and the corresponding upwards shift of the maximum values of R1. The resulting growth of the R1 concentrations at higher latitudes can also be followed and, at 33° lat., the gradual rise in the height of the maximum during the latter stages of the experiment which was produced by the horizontal eddy transport can be seen. The relatively minor increase in the R1 content at levels other than those where it was initially defined is also apparent.

3.3. TRANSPORT MECHANISMS

In the subsequent sections the details of the mechanisms by which R1 was diffused will be presented. For convenience the vertical and horizontal fields will be discussed separately, and then the total field will be presented in the form of the convergence. This analysis is based on zonal mean distributions time averaged over 10-day periods, 18 such distributions being available from this study. The standing eddies that exist in the actual atmosphere do not occur in the model because of the lack of topography. However, it is thought that the large-scale eddies of the model replace the role of both the transient and standing eddies of the atmosphere. It should be noted that the horizontal components of the large-scale eddy and mean meridional fluxes given here are multiplied by the mass of the atmosphere, 1000 gm./cm.^{2} Also, the horizontal component includes a term $2\pi a \cos\theta$ indicating that the flux has been integrated around a latitude circle.

It should also be pointed out that in the subsequent discussion the two meridional cells that existed in the model stratosphere will be referred to for convenience as the indirect cell (high latitude) and the direct cell (low latitude). In the thermal sense both of these cells are indirect, although there is some justification for considering the low latitude cell to be direct since it is an extension of the direct meridional cell in the troposphere.

3.3.1. *Vertical transport.*—In figure 6 the vertical fluxes of R1 owing to the meridional circulation and the large-scale eddies are compared for three 10-day periods which represent various features of the evolution of this tracer. The initial development given in figure 6a, b was limited to the low latitudes in accordance with the R1 distribution. In the Tropics the large-scale eddy diffusion was directed downgradient on either side of the level of maximum concentration, while the upward eddy flux in middle latitudes was countergradient at the lower levels and largely counteracted the downward branch of the meridional cell.

As R1 diffused polewards the magnitude of the tracer transported by the tropical meridional cell reduced, while transport of R1 by the indirect cell at higher latitudes increased, and by the end of the experiment this cell supplied the larger contribution to the R1 diffusion. The downward branches of the two cells reinforced one another resulting in a general descent of R1 over the latitude range 25° to 50°. The maximum intensity of the mean meridional fluxes was somewhat greater than that of the eddy fluxes throughout the experiment.

The vertical eddy fluxes changed remarkably during the course of the experiment, with the fluxes in the Tropics disappearing as the concentration gradient fell following the diffusion of R1. At the end of the experiment there was an upward eddy flux over most of the middle stratosphere, which was dominated by the cell in middle latitudes. On the other hand, the troposphere and lower stratosphere had essentially downward eddy fluxes, the maximum being situated in the region of the tropopause gap as figure 6f shows. The division of the atmosphere into two eddy flux regions seemed to correspond rather roughly to the line of maximum concentration of R1; in both regions the eddies were essentially directed towards reducing the concentration gradient of R1. After about 40 days these cells lay approximately on a line joining the two jet streams as figure 6f indicates, a similar orientation being observed for the eddy downward flux of R2. The location of the lower eddy cell in the region of the tropopause gap is controlled by the jet stream, since the eddy kinetic energy and presumably the vertical eddy velocities have maximum values in its vicinity, as is reasonable to expect. This mixing ratio cell is located above the actual jet stream, and therefore the maximum eddy kinetic energy, because of the vertical concentration gradient of the tracer. High tracer concentrations are not maintained in the lower part of the jet stream because of their rapid diffusion in the troposphere.

A point of great interest revealed in figure 6e,f is the fluxes in the region of the tropopause gap, as they suggest that the principal transport into the troposphere from the stratosphere occurs in this region. The eddies seemed to be most important for downward transport at the gap itself, the downward branch of the tropical meridional circulation then predominating for the transport into the troposphere. Apart from the subtropics there was a general, very weak, downward eddy flux in the troposphere and across the tropopause at all latitudes. There was no indication that the tropopause acts as a "barrier," as is sometimes stated, to the transport from the stratosphere to the troposphere.

3.3.2. *Horizontal transport.*—The horizontal fluxes were approximately three orders of magnitude greater than the vertical fluxes, in accordance with the relative magnitudes of the respective velocities. This however does not mean that the horizontal fluxes completely dominated the R1 transport, as will be seen from the following section when the flux convergences are discussed.

In figure 7, 10-day time means have been given for the same three periods as in the previous section, mean meridional and eddy fluxes again being presented separately. A general feature of the results revealed by this figure is that the mean meridional fluxes were not as dominant compared to the eddies as they were for the vertical transport. Also the meridional cell pattern was not as well defined as in the vertical case, and it was apparent for some 10-day means (not included in figure 7) that a time scale of 10 days was too small to filter out the noise. The mean meridional flux was strongest for the first 10-day period, even then being less than half the intensity of the corresponding eddies. After about 20 days this flux weakened and an equatorwards cell appeared (see fig. 7c) which was at level 5 but at 40° lat. This cell remained remarkably constant in height, latitude, and intensity and seemed at times to be a more permanent feature of the flow than the R1 transport associated with the tropical meridional cell. Most of the time the R1 flux of this midlatitude cell was weaker than that of the tropical cell. The R1 flux produced by the direct meridional cell in the Tropics fluctuated quite widely in intensity and form, at times appearing to consist of two components, one being centered at about 25°, the other at the Equator (see fig. 7c).

The horizontal eddies showed considerable development during the experiment, although not as much as the vertical eddies. The eddy flux had its maximum intensity during the first 10-day period, and then gradually reduced and extended over a wider latitudinal range as a comparison of figures 7b and d indicates. The large northward eddy cell fluctuated noticeably between 10-day means, and as in the case of the mean meridional flux appeared at times to have two centers of activity. The existence of these two centers can be seen in figure 7f, one center generally being at about 40°, the other at 20°. The 40° center was at level 6, presumably associated with the downward transport in this region, while the 20° center was at level 4 following the shift of the maximum concentration of R1 in the Tropics. After about 30 days a countergradient eddy cell, which transported R1 equatorwards,

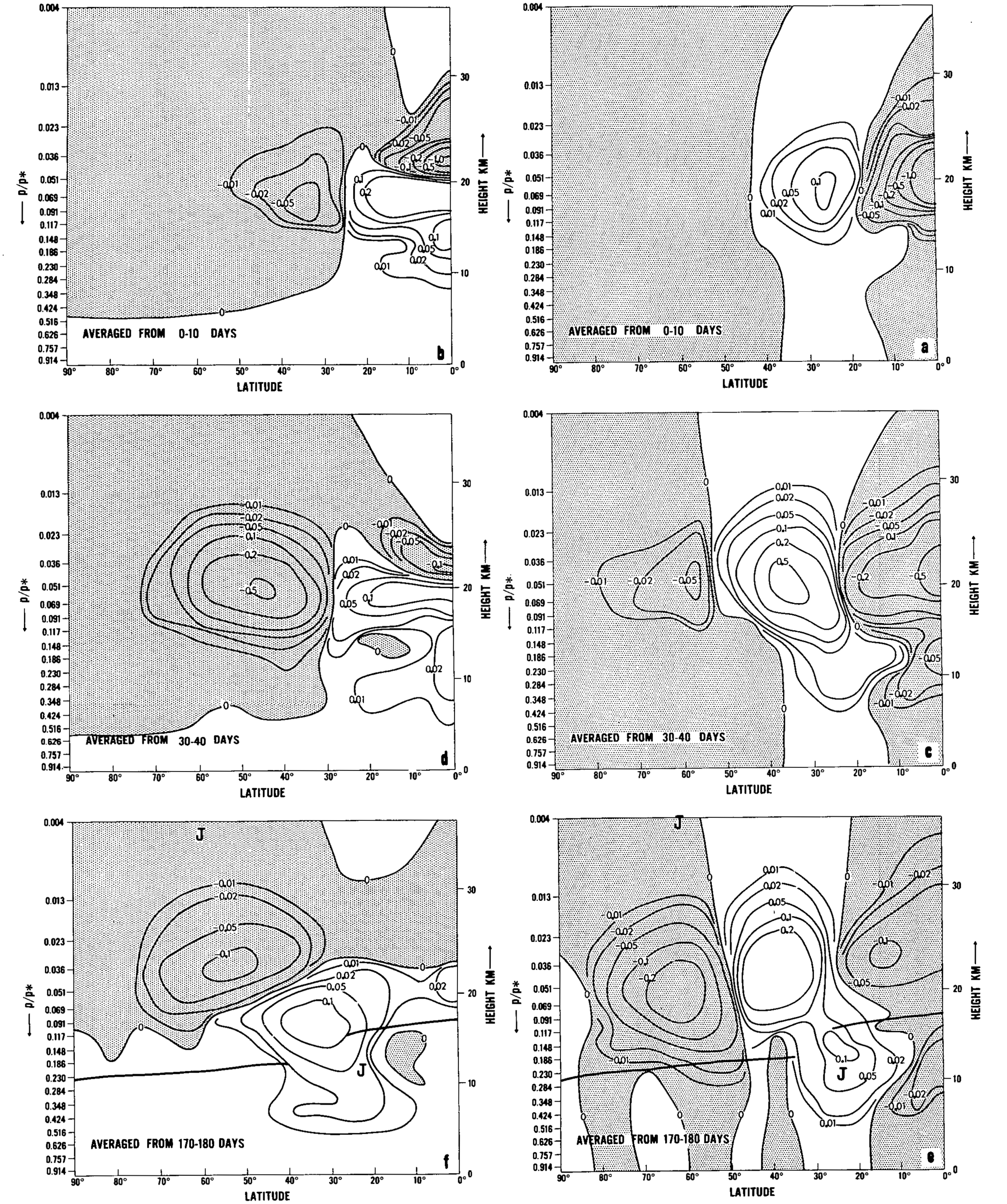

FIGURE 6.—Downward fluxes of R1 by the meridional circulation and large-scale eddies. (Units: $\frac{\text{gm.}}{(\text{cm.}^2\ \text{sec.})}\times 10^6$.) Eddies are on the left, the meridional circulation on the right. Shaded areas are upwards fluxes. The heavy black line indicates the approximate position of the tropopause and "J" indicates the positions of the jet streams.

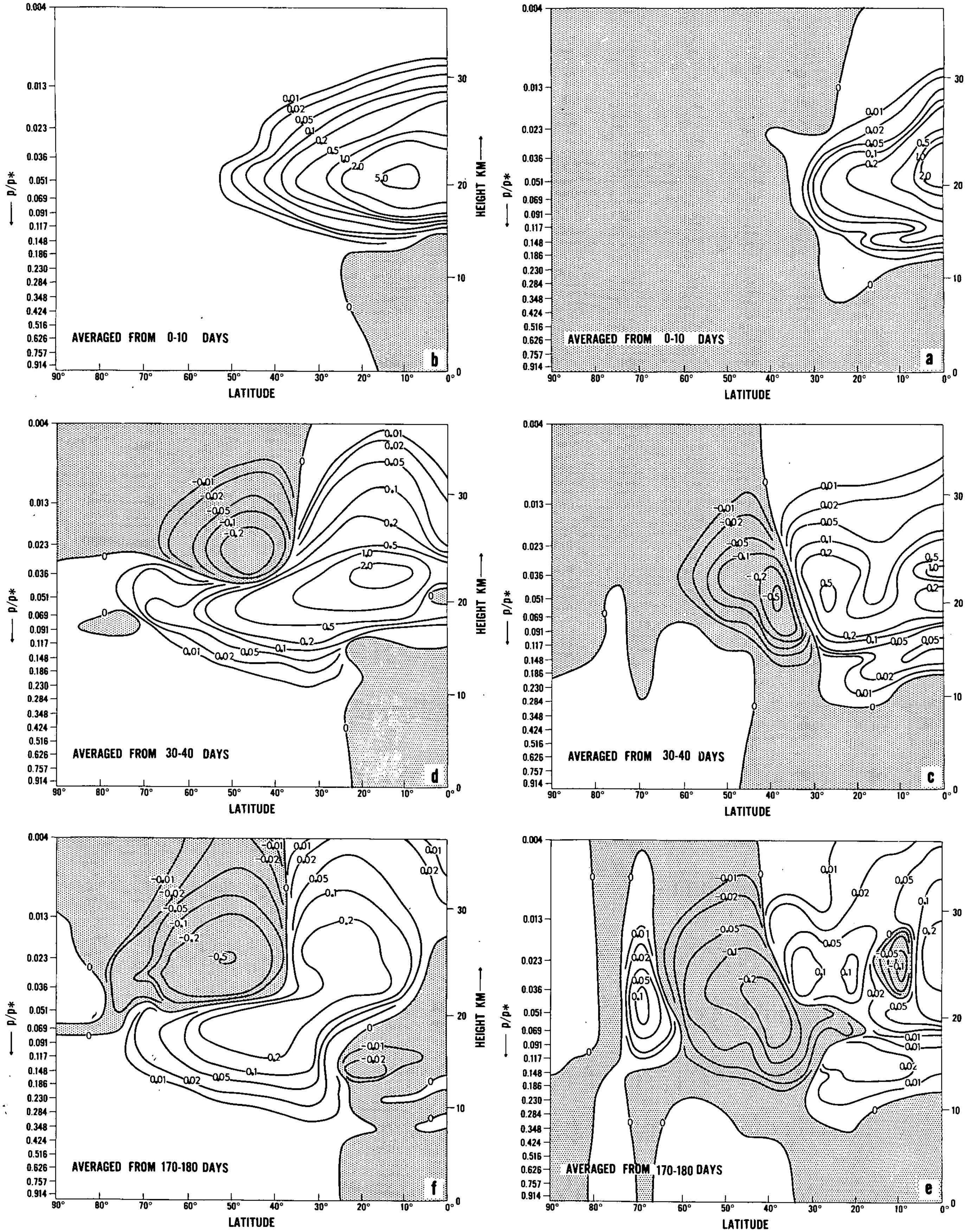

FIGURE 7.—Horizontal fluxes of R1 by the meridional circulation and large-scale eddies. (Units: $\frac{\text{gm.}}{\text{(atm. sec.)}}\times 10^{-13}$.) Eddies are on the left, the meridional circulation on the right. Shaded areas are equatorwards flow.

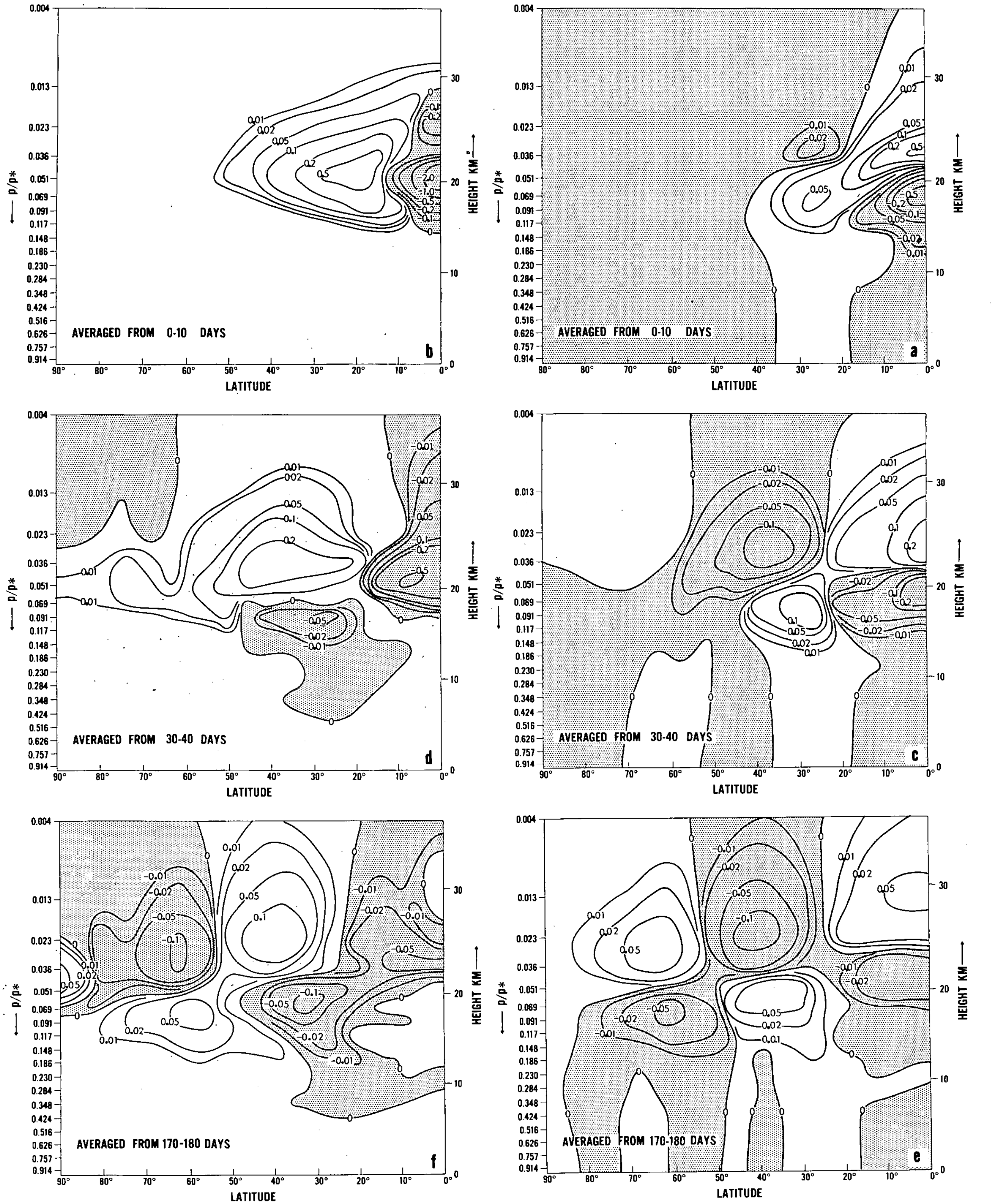

FIGURE 8.—Total rate of change of R1 by combined vertical and horizontal components. (Units: $\frac{\text{gm.}}{(\text{gm. sec.})}\times 10^7$.) Eddies are on the left, the meridional circulation on the right. Shaded areas are regions where R1 is decreasing.

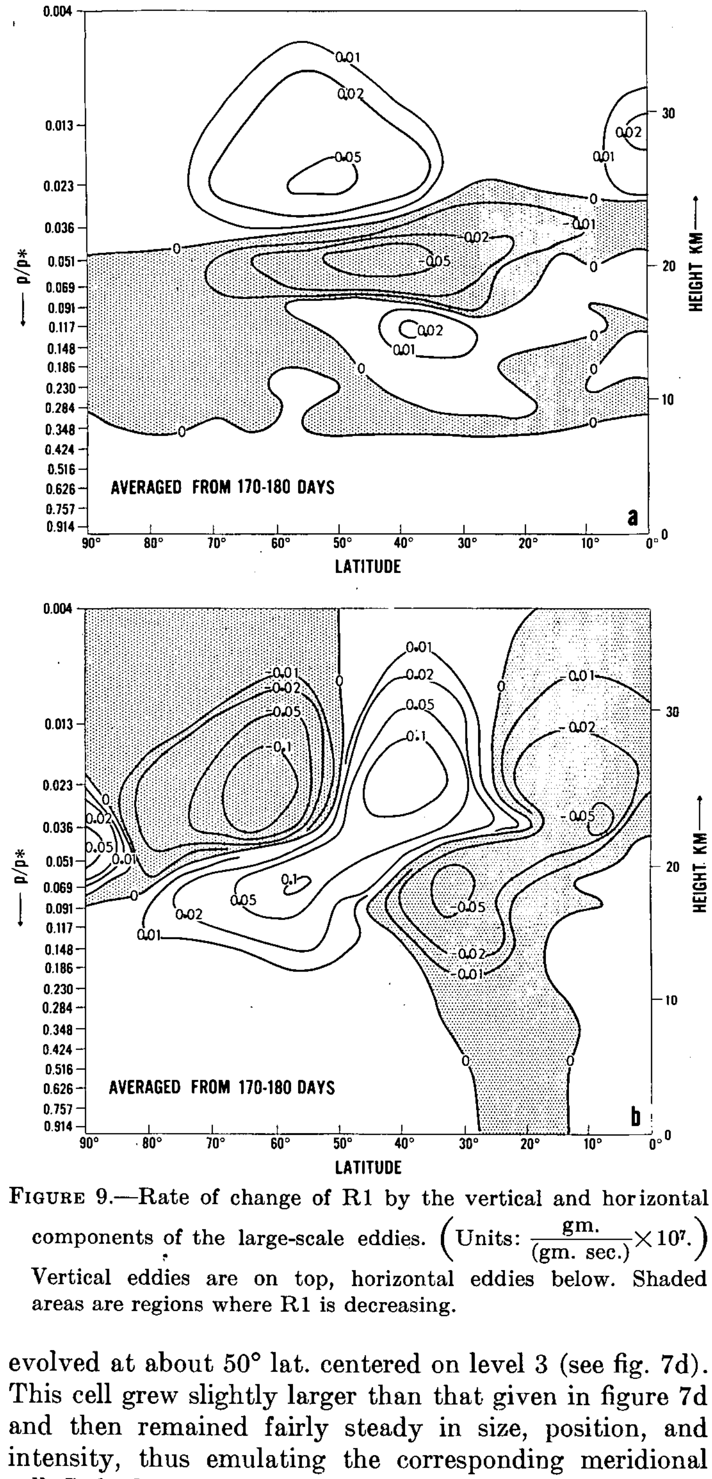

FIGURE 9.—Rate of change of R1 by the vertical and horizontal components of the large-scale eddies. $\left(\text{Units: } \frac{\text{gm.}}{(\text{gm. sec.})} \times 10^7.\right)$ Vertical eddies are on top, horizontal eddies below. Shaded areas are regions where R1 is decreasing.

evolved at about 50° lat. centered on level 3 (see fig. 7d). This cell grew slightly larger than that given in figure 7d and then remained fairly steady in size, position, and intensity, thus emulating the corresponding meridional cell. It had approximately half the maximum intensity of the much larger gradient eddy cell, except for the last 10 days when for the first time it exceeded it. There was also a very weak equatorwards flux below about level 7 in the Tropics, apart from this there was a steady northward eddy flux over the whole of the troposphere and lower stratosphere.

3.3.3. *Convergences*.—In figure 8 the total flux convergences (i.e., the combined vertical and horizontal components) for the eddies and the meridional circulation are compared for the same three 10-day averaged periods as in the previous two sections. In the case of the meridional circulation the analysis into vertical and horizontal convergences is not meaningful, since the divergence due to the horizontal component of the meridional circulation is almost completely compensated for by convergence of the associated vertical component. Vertical and horizontal eddy convergences were available, but for brevity only the values for the last 10-day period are given in figure 9; these can be compared with the total eddy convergence of figure 8f. On the whole the horizontal convergences were larger, being approximately twice the vertical convergences. An interesting result revealed by the breakdown given in figure 9 is that the two eddy components opposed one another in some regions of the atmosphere.

The combined convergences of figure 8 show that the initial values were the highest for the whole experiment. While the initial eddy flux mainly transported R1 from the Tropics to the subtropics with minor vertical transport, the meridional circulation primarily induced changes in the vertical distribution. Thus at the Equator R1 was transported upwards from levels 6 and 7 into 4, while around 30° lat. the reverse happened with divergence from level 4 and convergence principally into levels 6 and 7.

As R1 diffused, the interrelationship of the eddies and the meridional circulations was gradually revealed. An indication of this is given in figure 8c,d, where it can be seen that these two transport mechanisms were in opposition at most points, although at this early stage they were nowhere near in balance, and therefore the growth of R1 away from the Tropics continued at a high rate. The interesting feature in the eddies at this time is the weak divergent region centered in the lower stratosphere of the subtropics (fig. 8d) which grew in order to counteract the convergence there by the meridional circulation.

During the remainder of the experiment, the initial maximum eddy divergence at level 5 gradually weakened, while the secondary related center of divergence which developed around level 6 in the subtropics became more intense as the convergence by the meridional circulation in this region increased. Another eddy divergent region slowly developed in the middle stratosphere at high latitudes which opposed the meridional convergence there by the indirect cell. The essential features of the meridional circulation during this time was an improvement in the definition of the convergent pattern given in figure 8c, part of this being in response to the growth of the eddies. The divergent-convergent pattern in the subtropics remained fairly constant, despite the weakening of the contribution from the Tropics, because of the intensification of the contribution by the indirect circulation at high latitudes. This contribution became more prominent as R1 diffused and resulted in convergence in the upper polar stratosphere and divergence in the lower part. At the stage of the R1 diffusion given in figure 8e,f, it is apparent that neither the eddies nor the meridional circulation dominated the flow pattern, and their primary concern seemed to be to neutralize each other's contribution to the rate of change of R1.

Finally mention should be made of the contribution of the subgrid scale diffusion in the evolvement of the R1 distribution. In figure 10 this is shown for the last two of

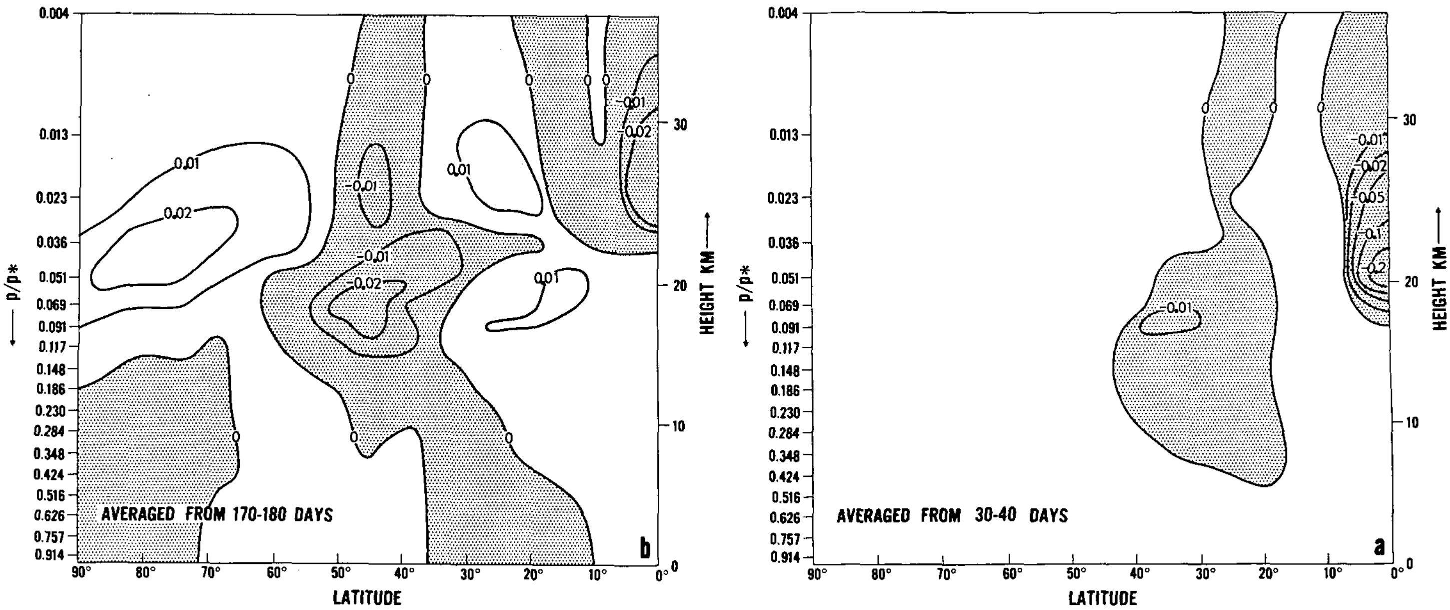

FIGURE 10.—Rate of change of R1 by the subgrid scale horizontal diffusion. $\left(\text{Units: } \frac{\text{gm.}}{(\text{gm. sec.})} \times 10^7.\right)$ Shaded areas are regions where R1 is decreasing.

the three 10-day periods; for the first 10-day period this form of diffusion was virtually zero except right at the Equator. Even for the fourth 10-day period (fig. 10a) the diffusion only really acted at the Equator. However between 40 and 80 days the R1 subgrid scale diffusion changed quite extensively, and subsequently remained fairly constant in its form and particularly in maximum intensity. As figure 10b shows the diffusion pattern grew into a series of positive and negative areas which were of quite important magnitude, having about 20 percent of the intensity of either the meridional circulation or the eddies.

In order to provide a clearer idea of the relative importance of the various transport mechanisms, these have been combined for the last 10-day period and are shown in figure 11 for two levels. The dominant feature at both levels is the near balance between the rates of change produced by the meridional circulation and the eddies. The net rate of change of R1 given in figure 11 was computed from the difference between the tracer concentrations at the beginning and end of the 10-day period, rather than as the residual of the various mechanisms. This gives a somewhat smoother distribution, as the errors involved in the convergence calculations are removed. The net rate of change is generally quite small, being essentially the residual between the eddy and mean meridional transports, and illustrates the rather poor efficiency of the atmosphere as regards the diffusion of R1 once the initial concentration gradients had been reduced. Between levels 3 and 6 the signs of the major convergences change, a result also apparent in figure 8e, f, while the magnitude of the convergences is also somewhat smaller at the lower level.

3.4. EXPLANATION OF THE R1 DIFFUSION

It is of considerable interest at this stage to collect together the results presented in the previous sections in an attempt to give a coherent account of how the R1 tracer was transported in the model atmosphere. As will be seen, the meridional circulation played an important part in the tracer diffusion, and it should be recalled that figure 9 of Part I shows that a 2-cell structure existed in the model stratosphere.

3.4.1. *Large-scale transport features.*—Because the R1 diffusion was somewhat involved, a description based on the schematic diagram in figure 12 will be given prior to discussing the transport mechanisms in more detail. This figure, although rather idealized, is essentially based on the convergence patterns given in figure 8 for the last 10-day period. In figure 12 the maxima of the regions of convergence by the large-scale eddies and meridional circulations are indicated by E^+ and M^+ respectively, these regions are associated with companion divergent regions, M^- and E^- respectively, which are omitted for clarity. The numbers in the figure give the approximate chronological order of events involved in the evolution of the tracer distribution, although events 1 and 2 actually occurred simultaneously. The figure shows the upward transport by the direct cell, event 1, which progressively removed R1 from the lower levels, and accumulated it in the higher levels of the Tropics. This largely accounts for the equatorial minimum obtained in level 5 and below and the corresponding equatorial maximum in the upper levels, details of which can be seen in figures 1 and 2. Concurrently the horizontal component of the large-scale eddies transferred tracer from the high concentration region where R1 was initially placed and deposited it in the subtropics, event 2, producing the E^+ region shown in the upper levels. The tracer did not accumulate there because the downward branch of the direct cell in the subtropics, event 3, transported this tracer from the E^+ to the M^+ region in the vicinity of the tropopause gap. From this M^+ region the large-scale eddies removed R1 quasi-horizontally, event 4, primarily polewards in the

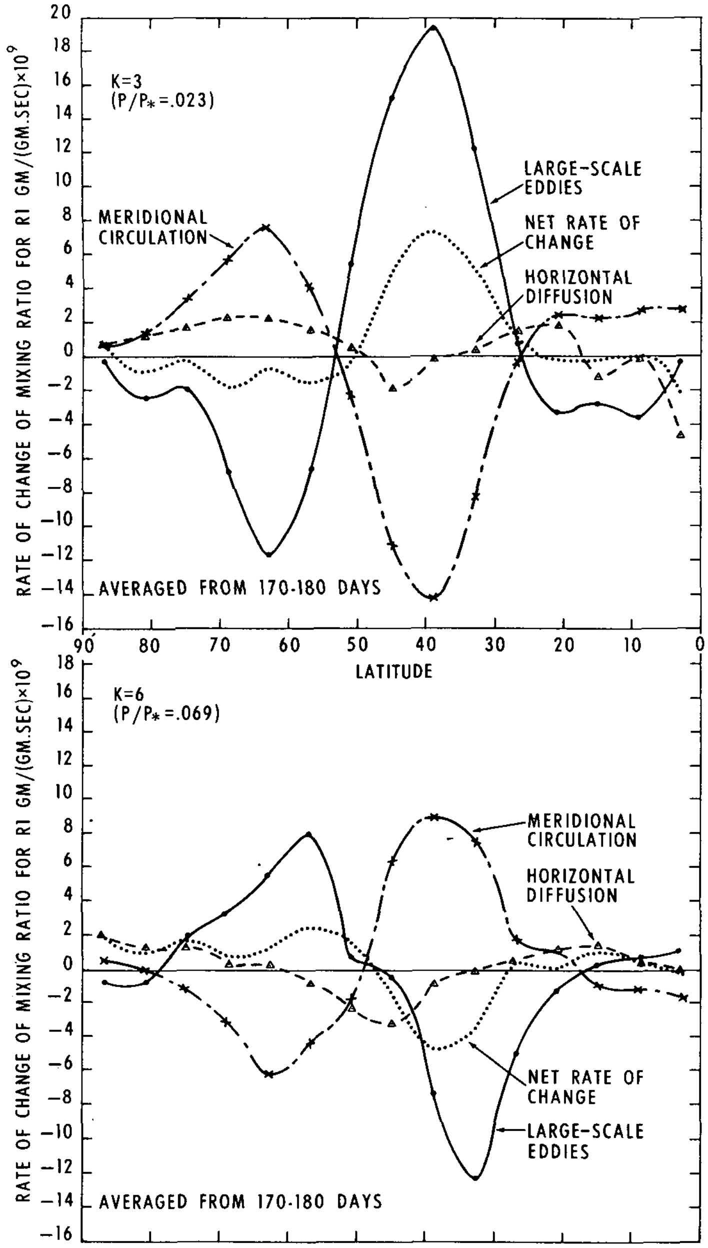

FIGURE 11.—Rate of change of R1 versus latitude for the various transport mechanisms averaged over the last 10 days of the experiment. Results are shown for levels 3 and 6. $\left(\text{Units:}\frac{\text{gm.}}{(\text{gm. sec.})}\times 10^9.\right)$

lower stratosphere producing the E^+ region at high latitudes. Some polewards eddy transport also occurred in the upper troposphere which removed most of the R1 entering the troposphere in the vicinity of the tropopause gap. A much weaker R1 eddy transfer took place from the subtropics to the Equator which helped to replace the tracer removed from that region by the upwards branch of the direct cell. The tracer from the lower stratosphere was then transferred to the middle polar stratosphere by the upward branch of the indirect cell, event 5, accounting for the M^+ region there. This in turn supplied tracer to the horizontal eddies which advected it to midlatitudes, event 6, thus producing a joint convergence region with that associated with the horizontal fluxes of event 2.

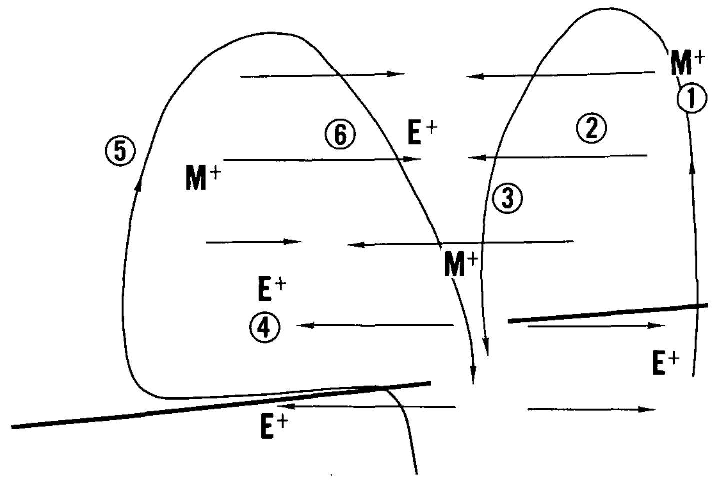

FIGURE 12.—Schematic diagram showing the combined actions of the large-scale eddies and the meridional circulation for quasi-steady state conditions for R1: E^+ and M^+ represent regions of convergence by the eddies and meridional circulation respectively. The numbers indicate the approximate chronological order of events, and the horizontal arrows the horizontal eddies.

Finally the return branch of the indirect cell transported downwards into the subtropical M^+ region the tracer converged into the midlatitudes by event 6, thus completing the cycle. The important result revealed here is the interaction and mutual opposition of the tracer transports by the large-scale eddies and the meridional cells, which results in a rather slow diffusion of the tracer.

Figure 8d reveals a feature which is missing from the schematic diagram of figure 12. This is the extensive eddy convergence zone centered on about level 5, which was produced by a direct eddy transfer of R1 from the Equator to the Pole as shown in figure 7d. This direct transfer was responsible for most of the R1 which accumulated in the higher latitudes at these levels during the earlier part of the experiment; subsequently transport from the lower stratosphere via the upwards branch of the indirect cell became of increasing importance. This direct transfer mechanism fluctuated considerably between 10-day mean periods, and after about 60 days at level 5 and above it did not generally extend polewards of about 60° lat. At levels 6 and 7 there was frequently a direct Equator to Pole transfer, but most of the flux originated from the meridional convergence zone in the subtropics.

3.4.2. *Latitude-height development.*—Considering now the latitude-height distributions of R1 shown in figure 4, it can be seen that they agree with the scheme given above for the diffusion processes. The initial development was essentially horizontal because of the dominance of the horizontal component of the large-scale eddies, but there was some upwards movement in the Tropics which was associated with the direct cell. There followed a gradual development of a region of high concentration in the subtropics, largely produced by the downward branch of the direct cell in the early stages, which resulted in the zonal rings given in figure 2. This enabled the R1 concentration to be built up in the lower stratosphere at high latitudes, and this increase was considerably faster than that occurring at higher levels at these latitudes as figure

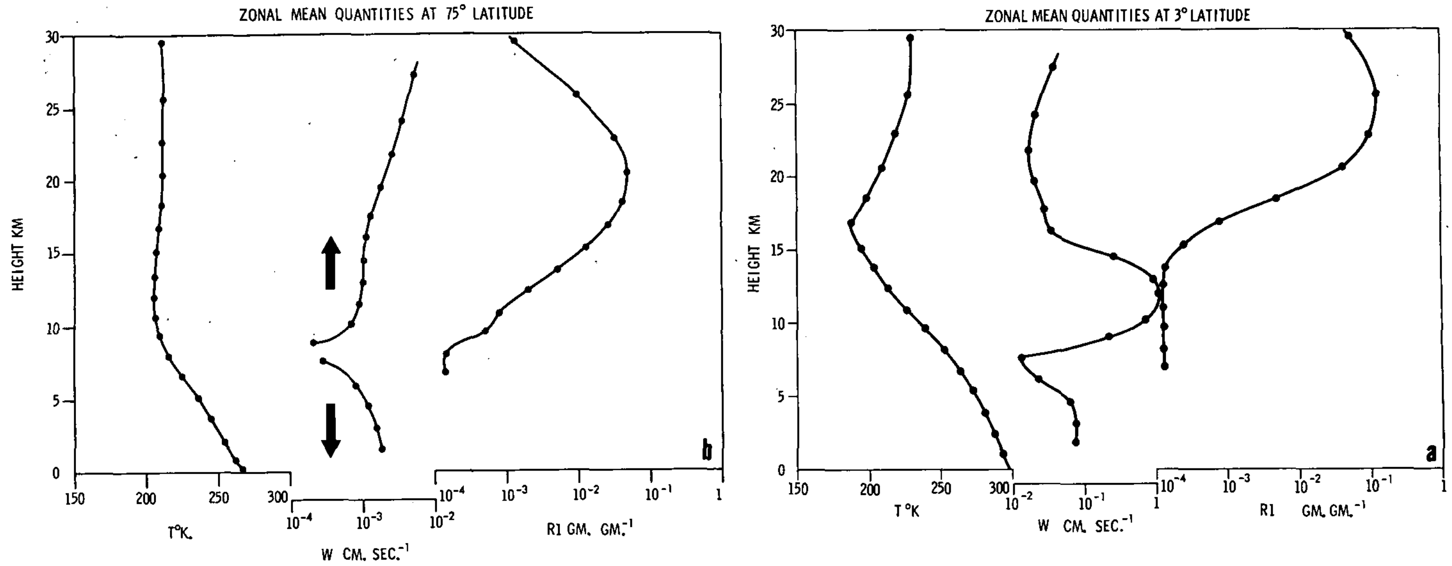

FIGURE 13.—Comparison of zonally averaged vertical velocity, temperature, and mixing ratio for R1 at 3° and 75° lat. The temperature and vertical velocity are 100-day mean quantities, while the mixing ratios are instantaneous values for the last day of the experiment. Note the different horizontal scale for vertical velocity at the two latitudes.

4e shows. The reason for this can be attributed to the horizontal components of both the large-scale eddies and the indirect meridional cell transferring the tracer equatorwards, and thus countergradient, in the middle polar stratosphere. This meant that this region had to be supplied from the lower levels which made it the last link in the transport chain. The final mixing ratios in the lower stratosphere at extratropical latitudes showed little variation with latitude, indicating that the horizontal component of the eddies dominated the R1 concentrations in this region.

The final R1 distribution in figure 4f reveals that in the middle stratosphere there was a general downwards slope of the isopleths from the Equator to the Pole, although the figure very greatly exaggerates the actual slope because of the scale used. This slope arose partially because the source of R1 was at the Equator and the maximum concentrations always remained there, while steadily rising in altitude because of the upwards flow produced by the direct cell. Thus, even if the tracer had been able to travel directly to the Pole horizontally, this upwards movement would have resulted in some slope appearing. In addition, since the tracer supply for the high latitudes is actually transferred downwards in the subtropics by the meridional circulation and vertical eddies before it is taken polewards by the horizontal eddies, this tends to accentuate the slope. This combination of upwards and downwards tracer fluxes in the Tropics and subtropics, respectively, also accounts for the basic shape of the R1 distribution in the vicinity of the tropopause, the direct meridional cell being of major importance in determining this shape.

The movement of the maximum R1 concentration in the Tropics is of interest, since it provides an indication of the influence of the meridional circulation on the tracer distribution, which will be compared with observation in a later section. The initial maximum moved from level 5, or about 20.5 km., to a position between levels 3 and 4 of approximately 24 km. 180 days later. Since the zonal mean vertical velocity in the Tropics (3°) time averaged over 100 days was about 2×10^{-2} cm. sec.$^{-1}$ in this altitude range, in a time interval of 180 days a particle would have moved upwards 3.1 km. This compares favorably with the value of 3.5 km. estimated from the R1 distribution in figure 4, and the tropical meridional cell can therefore adequately account for the upwards displacement of the R1 maximum. As far as the latitudinal variation is concerned the maximum R1 concentration does not appear to have moved from the Equator during the course of the experiment. The 100-day mean meridional velocity at level 5 was approximately 3 cm. sec.$^{-1}$ in the Tropics (3°) and was directed polewards. (The corresponding velocities at levels 4 and 6 were about double this at 3° lat. but were directed equatorwards.) Over a period of 180 days the level 5 velocity would therefore have moved a particle about 470 km. polewards, but since the model grid size at the Equator is 320 km. this distance only amounts to transfer between adjacent grid points, making it barely discernible. In any case as the R1 maximum moved up to level 4 during the experiment the equatorwards velocity at this level would have helped to maintain the maximum at the Equator.

3.4.3. *Vertical distribution of* R1.—In the atmosphere it is observed that water vapor, ozone, and radioactive debris tend to have a discontinuity in their vertical profiles at the tropopause and, as discussed in Part I, in the case of water vapor and ozone the change in their concentrations at the tropopause greatly influences the stratospheric temperature structure. It is therefore a matter of some importance to account for the vertical profile of R1, as this should provide some enlightenment on the problem of tracer profiles in general.

Part of this problem consists of explaining how the very low R1 concentrations were maintained in the troposphere

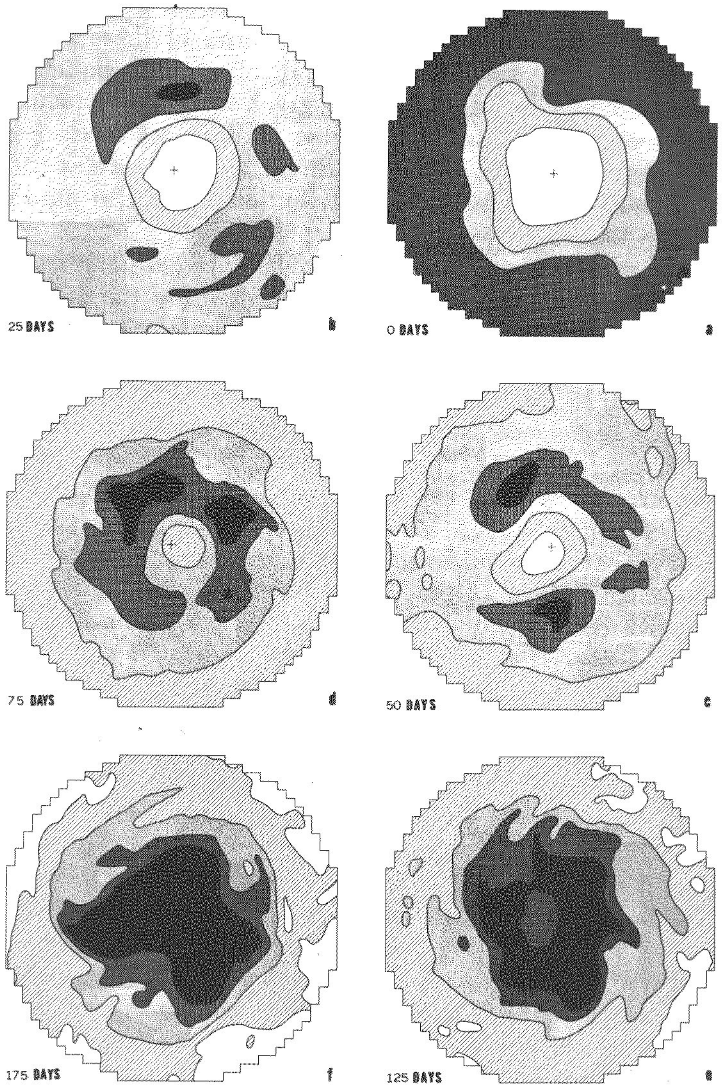

FIGURE 14.—Hemispheric variation of R2 at level 4 for various instants of time. (Units: μ gm./gm.) Coding used for mixing ratio concentrations are: 0–0.3, white; 0.3–0.7, hatching; 0.7–1.1, light gray; 1.1–1.5, dark gray; greater than 1.5, black.

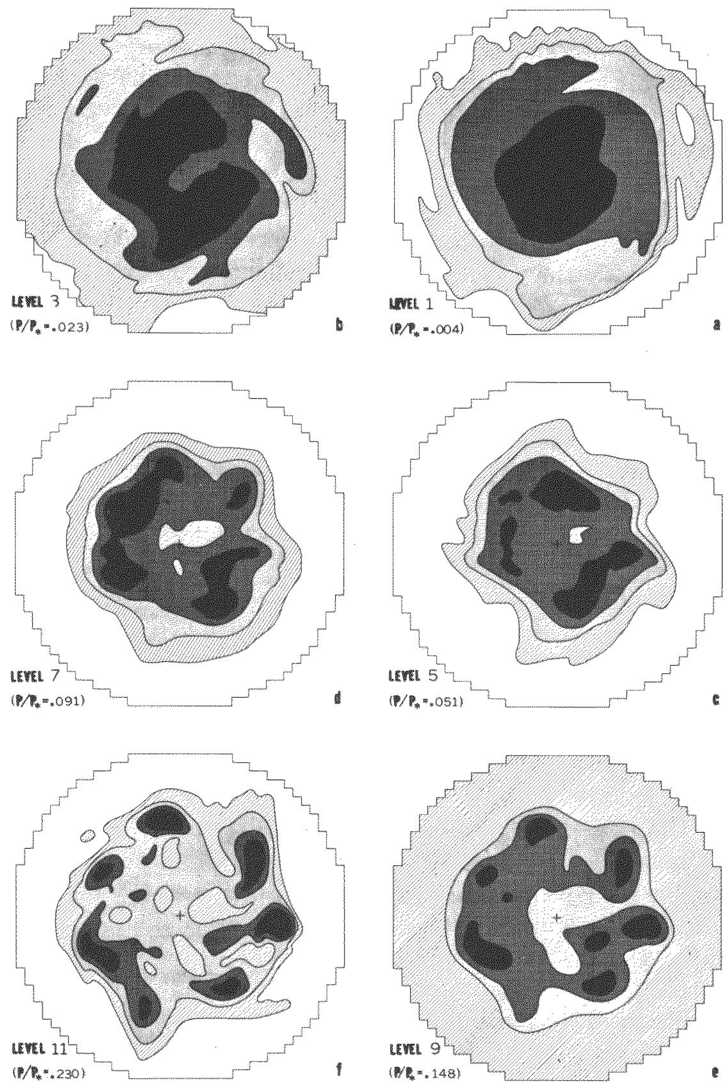

FIGURE 15.—Hemispheric variation of R2 for various levels 150 days after initiation. The shade coding represents different ranges of values of R2 for each level, so that individual values are not given. In each case the shading is used consistently with black representing the highest and white the lowest values.

throughout the experiment. The principal source of R1 for the troposphere was the downwards flux by the vertical eddies, and particularly the meridional circulation, through the tropopause gap as revealed in figure 6e,f. Now, since this downwards flux was continually reduced by the polewards transport of R1 by the large-scale eddies in the lower stratosphere, only the residual flux actually penetrated to the troposphere, and it was this flux which had to maintain the tracer concentrations at all latitudes. However, the tracer content of the troposphere is not conserved, due to precipitation processes which are enhanced by the faster vertical mixing in the troposphere, and this together with the rather limited tracer supply results in the low tracer concentrations obtained. On the other hand, as indicated in figure 12, the stratospheric motions form an almost closed system as they cycle the tracer around, and this tends to conserve the tracer thus permitting much bigger equilibrium concentrations to be retained.

To complete the discussion specific R1 profiles at two latitudes will be considered. In figure 13 instantaneous R1 distributions for the last day of the experiment at tropical and polar latitudes are compared with their respective temperatures and vertical velocities time averaged over 100 days. (The kink in the tropical vertical velocity profile at 7 km. is associated with the reverse cell, the model generated in this region; see Part I.)

Dealing with the Tropics first it can be seen that the R1 concentrations in the troposphere were remarkably uniform with height, except that they started to increase somewhat below, rather than at, the tropopause. The stratospheric concentrations are essentially the residual of the original R1 distribution and will not be discussed here. The constancy of the R1 values in the troposphere appears to be maintained by the vertical eddies, which smoothed out differences in the R1 concentrations arising from the tracer transported by the horizontal eddies from the subtropics. This can occur in the troposphere, as opposed to the stratosphere, because of the greater efficiency of vertical mixing in the troposphere permitted by its lower static stability. We consider now the vertical velocities; figure 13a shows that the maximum velocity was at about 12 km., while that at the tropopause was a factor of 40 lower, giving a very high gradient of vertical velocity just below the tropopause. This reveals that most of the transport associated with the direct tropical cell was confined to the troposphere, as indicated in figure 9 of Part I. The crowding of the isopleths around the tropopause in figure 4 can then be attributed to the reduction in the upwards velocity as the tropopause was approached, which meant that isopleths at higher levels were not pushed up as much as those below.

In the case of the polar regions it is necessary to account for the stratospheric profile also. The maximum at level 5 in figure 13b appears to be related to the direct transfer of R1 by the horizontal eddies from the equatorial source, which occurred in the earlier stages of the diffusion (see fig. 7b,d). Because of the reduction mentioned previously of the downwards tracer flux in the subtropics, the tracer supply available for transport to higher latitudes by the horizontal eddies in the lower stratosphere decreased correspondingly with decreasing altitude. Hence there was an associated variation with altitude of the convergence by the large-scale eddies north of about 40° lat., as shown in figure 8f. This resulted in the steady decrease observed in the R1 concentrations in figure 13b below the level of maximum concentration. However, given sufficient time the R1 maximum would probably have moved to lower levels, as the highest convergence by the large-scale eddies occurred below the maximum at these latitudes. The remarkably uniform R1 latitudinal distribution, which existed north of about 40° in both the lower stratosphere and upper troposphere, can be attributed to the action of the large-scale eddies. However, a discontinuity exists in the vertical profile at the tropopause in this latitudinal region, an indication of which is given in figure 13b. The tropospheric concentrations also appear to be constant, although insufficient data were available to decide this, and presumably this was due to the action of the vertical eddies in the troposphere. Although figure 6f shows that there were downward eddies across the tropopause at high latitudes, the discontinuity in the R1 profile reveals their ineffectiveness in transporting tracer from the stratosphere, which leaves little doubt that horizontal transport from the region of the tropopause gap was responsible for maintaining the tracer concentrations in the upper troposphere.

4. DIFFUSION OF THE OZONE TRACER, R2

4.1. GENERAL DESCRIPTION

In discussing R2 basically the same procedure and order will be used in presenting the results as those for R1. The present section will therefore be devoted to describing in a very general way how R2 varied with time at the different vertical levels of the model. This discussion is based on a series of hemispheric maps of the instantaneous R2 distributions for these levels which were made at intervals of 25 days. Two series of these maps are given here; the first consisting of a time sequence for 6 different days for level 4 is given in figure 14, while the second consisting of a number of different levels for one particular day is shown in figure 15. A glance at the zonal mean latitude-height distributions of R2 in figure 16 may also be convenient at this stage.

A comparison of these R2 hemispheric maps with the corresponding maps for R1 demonstrates that the dispersion of a material in the stratosphere is greatly influenced by its concentration gradient. The R2 distribution was, of course, much more uniform than that of R1 and its latitudinal concentration gradient was always rather small, especially for the top few levels. The principal concentration gradient of R2 was in the vertical, and, unlike R1, was of constant sign, increasing upwards at all times. Because of these differences, the diffusion of R2 was dissimilar in many respects from that of R1 in appearance. The R2 concentration field did not undergo the drastic changes of R1, and the time scale associated with the R2 diffusion also seemed to be longer than that for R1.

On a hemispheric basis there was comparatively little change in the vertical distribution of R2, unlike R1, and at all times the top level was overwhelmingly dominant, having approximately three times as much tracer as all the other levels put together. Both levels 1 and 2 showed small losses in their total concentrations during the course of the experiment, of the order of 3 and 7 percent, respectively, while all lower levels made gains. At the lower levels where the initial concentrations were very small or nonexistent, quite spectacular increases occurred, but the actual amount of tracer involved was always extremely small. Therefore, the net effect of the diffusion for the top few levels was essentially to redistribute the tracer in the horizontal, while at the lower levels the changes were largely influenced by the influx of tracer into these levels.

The evolution of the R2 diffusion at the various levels can be discussed by referring to the results for level 4 in figure 14, since this level was fairly representative of all levels except the top. The initial tracer distribution in figure 14a was not completely symmetrical because the mixing ratio was calculated using the local pressure and temperature values at each point, and thus the tracer pattern reflects the asymmetries of these quantities. The top eight levels had a similar initial distribution with the maximum at the Equator and the minimum at the Pole, following the photochemical ozone values used. Mainly owing to the action of the upward branch of the meridional circulation in the Tropics the equatorial concentrations were gradually and continuously reduced at all levels and, except for level 1, replaced in the early stages by patches of high concentration in middle latitudes as figure 14b shows. These concentrations increased slowly and eventually a polar cap of high intensity was built up, at the same time the equatorial concentrations were being reduced to the minimum values for those levels. Thus the initial latitudinal concentration was

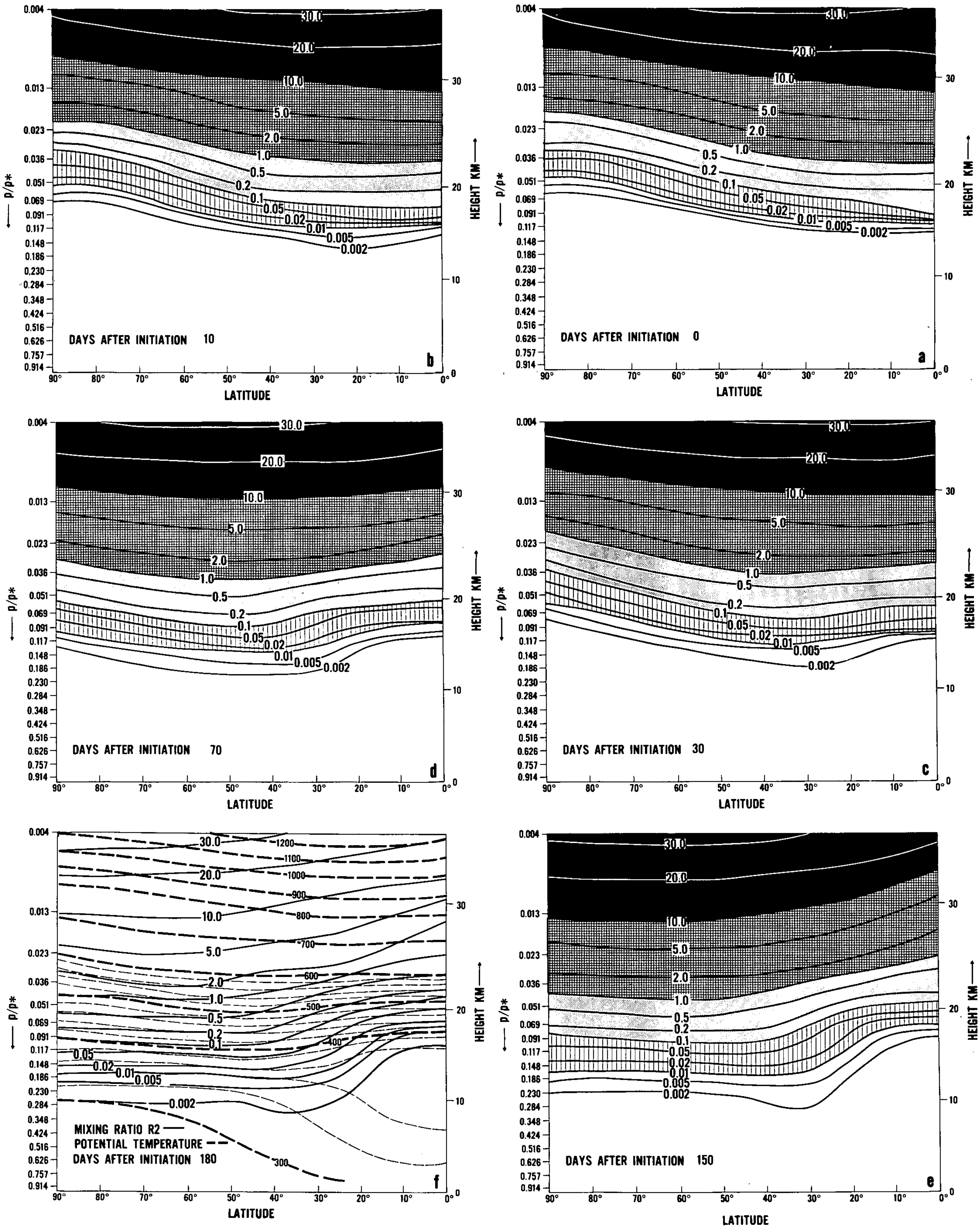

FIGURE 16.—Latitude-height distributions of the zonal mean R2 concentrations for various times. (Units: μ gm./gm.)

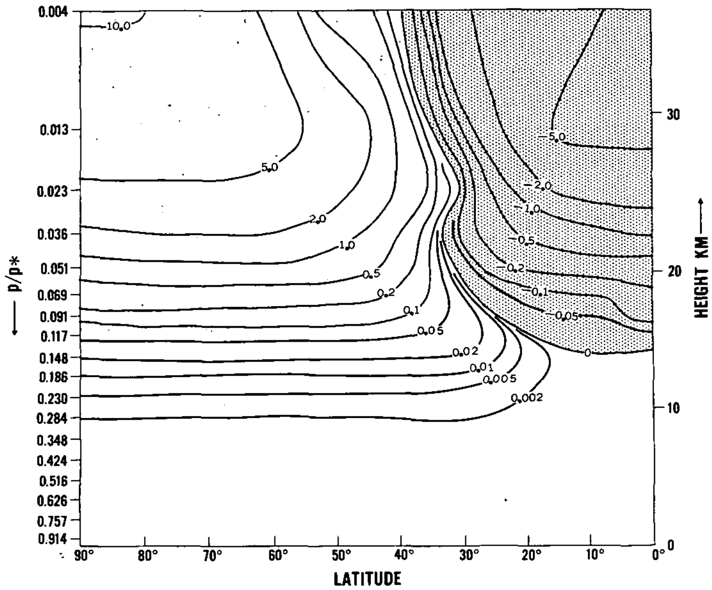

FIGURE 17.—Latitude-height distribution of the difference between the final and initial zonal mean R2 concentrations, illustrating the net changes produced in the R2 distribution by the dynamics. (Units: μ gm./gm.)

completely reversed. At no level was the maximum concentration actually at the Pole, but rather at a latitude which was increasingly closer to the Equator the lower the level considered, an indication of this variation being given in figure 15.

The sequence of maps given in figure 15 shows that, as for R1, there were clear correlations in the vertical between features of high concentration in adjacent levels. The correlations were actually much better defined for R2 than R1, and quite obviously became more closely related for the lower levels. Because of the high vertical concentration gradient of R2, downward transport from the higher levels was the dominating mechanism maintaining these correlations, the effect of transport from lower levels being unimportant. At the lower levels a distinct wave pattern could be distinguished in the tracer distributions, wave number 6 being shown in figure 15 at levels 7, 9, and 11. The tracer content of these lower levels was still growing strongly at the end of the experiment.

It can be seen from figures 14 and 15 that the lower latitudinal concentration gradient existing initially for R2 in no way seems to have suppressed the development of perturbations in the concentration with longitude. Although in general less marked than for R1, it is clear that these perturbations are a permanent feature of the flow, and as such are very important since they effectively represent part of the eddy transport.

4.2. LATITUDE-HEIGHT DISTRIBUTION

In figure 16 the time evolution of the zonal mean latitude-height distribution of R2 is given for the same "doubling" sequence of days as R1. The equatorial maximum concentrations, the small latitudinal and dominating vertical concentration gradients are all clearly shown in figure 16a. The dramatic feature of the time evolution is the reversal of the latitudinal concentration gradient from its initial photochemical conditions. This reversal occurred at all heights, and there is no doubt that the minimum concentrations at the end of the experiment were at the Equator. The percentage changes were smallest at the upper levels, and the development also seems to have been slowest there, although the fluxes and convergences were largest at these levels. After about 70 days the R2 latitudinal gradient was at its lowest value in the upper levels, and an examination of the horizontal eddy fluxes and convergences time averaged over a 10-day period around this time revealed that they also went through a noticeable minimum. The eddy fluxes still existed at this time even though the zonal mean R2 gradient was virtually zero.

At the lower levels the early development of the R2 transport was characterized by the appearance of local maxima in the subtropics. This was subsequently followed by a gradual accumulation of the tracer in the lower stratosphere at high latitudes, which went in phase with the R2 variation in the upper levels. At the end of the experiment there was scarcely any variation of the zonal mean R2 concentration with latitude in the lower stratosphere at middle and high latitudes. As the lower stratosphere received tracer the tropical troposphere lost it, and the isopleths in this region were gradually pushed up over the tropopause, resulting in a final R2 distribution in the lower stratosphere very similar in appearance to that for R1.

In order to illustrate the exact changes which occurred as a result of the diffusion the difference between the zonal mean R2 concentrations for final and initial conditions is displayed in figure 17. This shows that from the Equator to about 35° lat. R2 has been transferred to higher latitudes, although at the lower levels the latitudinal extent of the donor area is more restricted. The biggest changes were in the upper levels, and the R2 increases in the polar latitudes are particularly large. However, the most important feature is the increase in the tracer in the lower stratosphere below the levels where R2 was placed initially, which confirms that the dynamics enable the lower stratosphere to be used as a reservoir for suitable tracers. The zero line in this figure probably approximately represents the limit for annual mean conditions of the ozone source region of the atmosphere, this limit being dynamically rather than photochemically defined.

4.3. TRANSPORT MECHANISMS

As for R1, 18 sets of zonal mean distributions time averaged over 10-day periods were also available. Because R2 was initially distributed over a wider latitude and height range than R1 and in general had smaller concentration gradients, its fluxes and convergences were much steadier and therefore figures will be given for only the first and last 10-day periods.

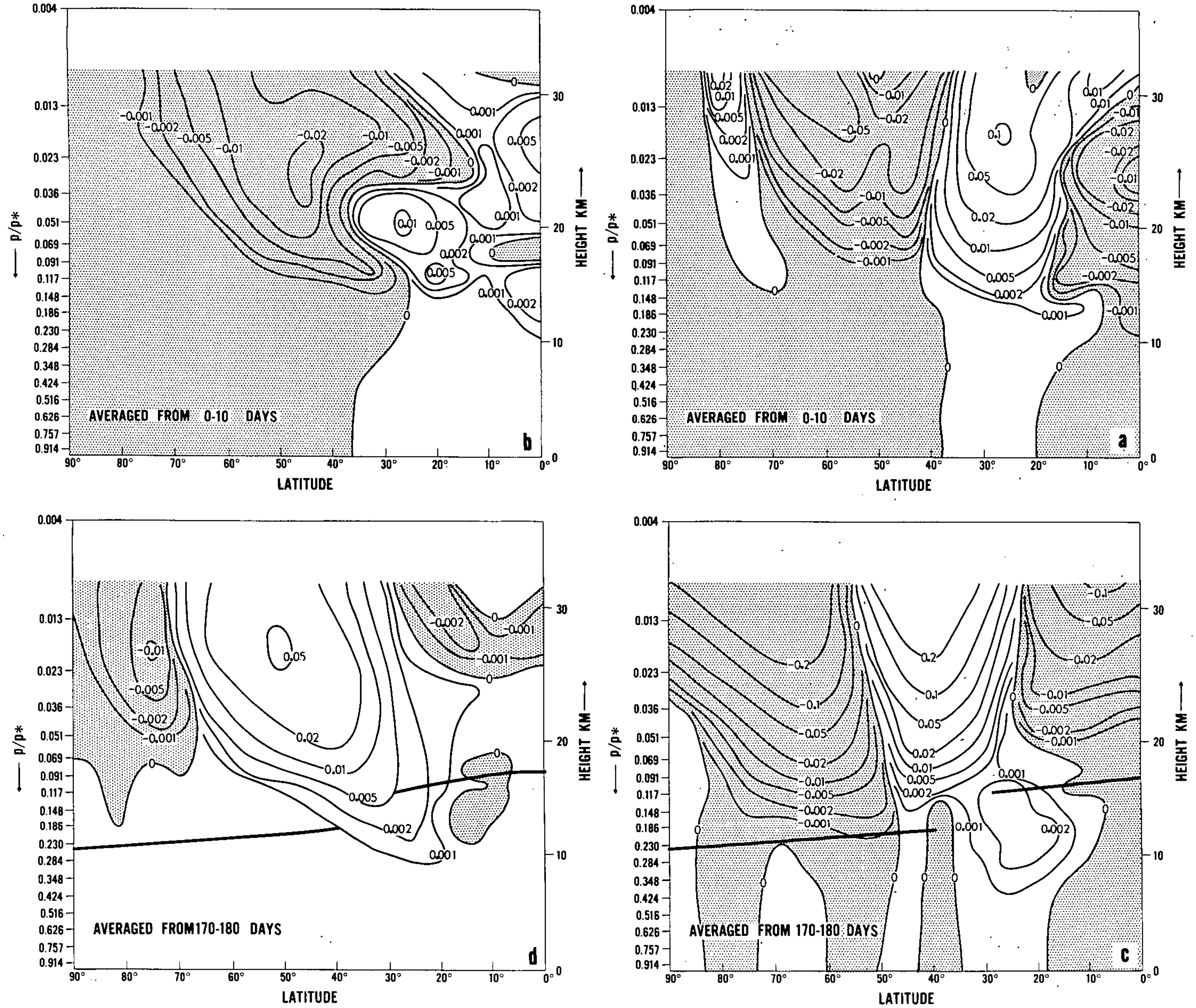

FIGURE 18.—Downward fluxes of R2 by the meridional circulation and large-scale eddies. $\left(\text{Units: } \frac{\mu \text{ gm.}}{(\text{cm.}^2 \text{ sec.})} \times 10^4.\right)$ Eddies are on the left, the meridional circulation on the right. Shaded areas are upwards fluxes. The heavy black line indicates the approximate position of the tropopause.

4.3.1. *Vertical transport.*—In figure 18 the vertical fluxes of R2 produced by the meridional circulations and the large-scale eddies are compared for the two 10-day periods. Several general features can be seen from this figure. Unlike the situation for R1, transport processes occurred at all latitudes immediately because of the initial R2 distribution, and the dominant fluxes were at all times confined to the upper levels of the model where the vertical gradient of R2 was maximum. The two-cell meridional flux pattern is clearly shown, with upward motion in the Tropics and high latitudes and downward motion in the subtropics. The maximum mean meridional fluxes were considerably larger than the corresponding vertical eddy fluxes.

The mean meridional fluxes were weakest during the first 10 days, with the fluxes in the Tropics being smaller than those at high latitudes; the latter were a consistent feature throughout the experiment. After about 30 days the meridional flux intensities increased quite sharply to values typical of those given in figure 18c for the last 10 days, and thereafter remained fairly constant. The cell pattern however varied more noticeably with the area of upward fluxes at high latitudes contracting somewhat. Also for some 10-day periods (see for example fig. 18a), a three-cell system appeared in the stratosphere owing to the development of a rather weak direct cell near the Pole. As for R1 the tracer transport associated with the downward branch of the tropical meridional cell penetrated into the troposphere in the region of the tropopause gap.

The vertical transport by the large-scale eddies developed more noticeably than that by the meridional circulation, as a comparison of the figures shows. The maximum

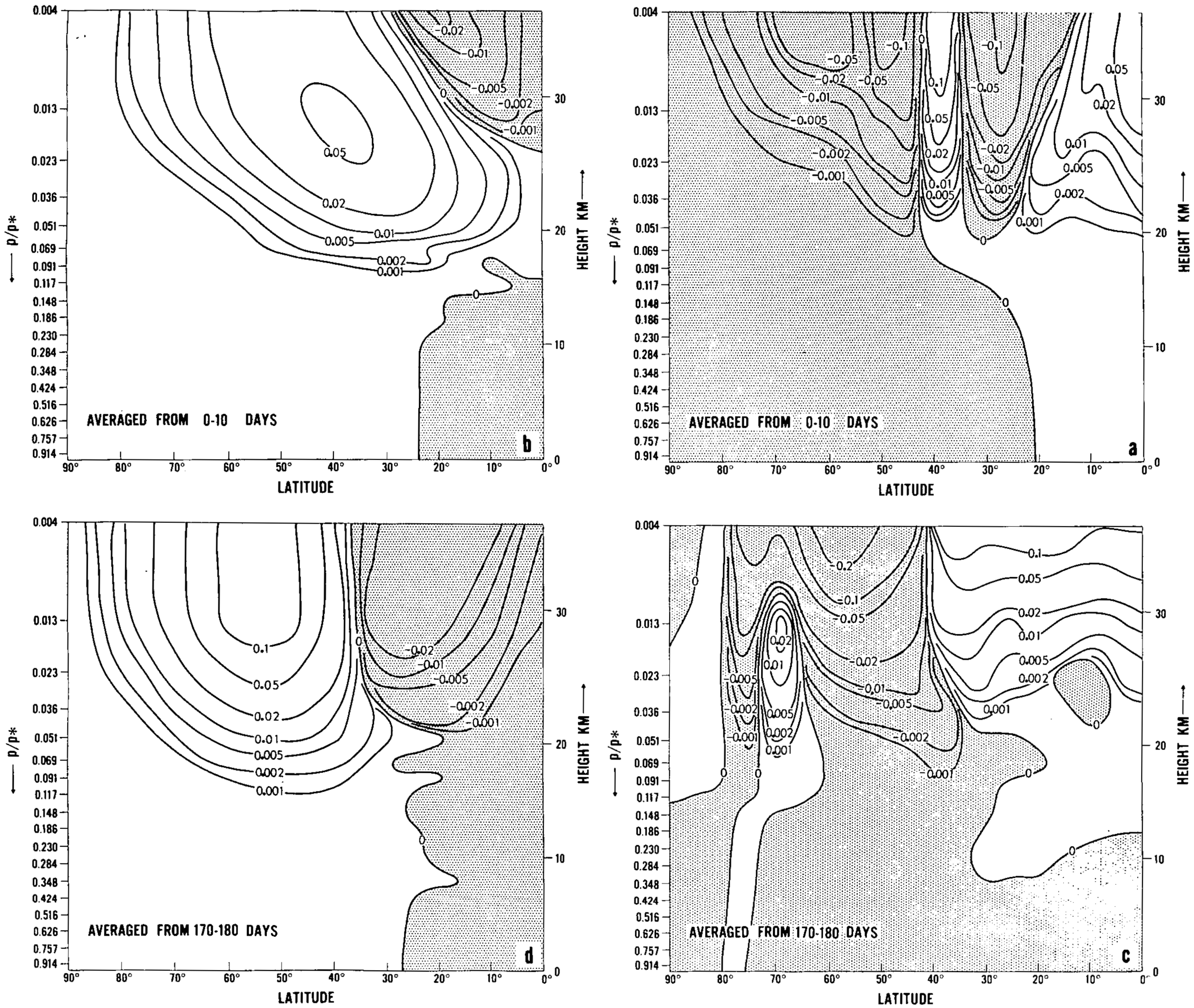

FIGURE 19.—Horizontal fluxes of R2 by the meridional circulation and large-scale eddies. $\left(\text{Units: } \frac{\mu \text{ gm.}}{(\text{atm. sec.})} \times 10^{-16}.\right)$ Eddies are on the left, the meridional circulation on the right. Shaded areas are equatorwards flow.

intensity of both the upward and downward eddies fluctuated by a factor of about five during the experiment, with those for the last 10 days being the most intense. The small downward eddy cell in the subtropics shown in the first 10 days gradually enlarged as the R2 was transported, first expanding into the Tropics and then moving more slowly upwards and polewards. It became firmly established after about 50 to 60 days and subsequently dominated the vertical transport by the eddies, generally having a maximum intensity greater than that associated with the upwards transport by the vertical eddies. This downward eddy cell was approximately aligned along a line joining the two jet streams (cf. R1), and there are indications that in the region of the tropopause gap the eddies are supporting the meridional circulation in transporting R2 into the troposphere. From the subtropics to the Pole there was downwards transport by the eddies at all latitudes in the lower stratosphere and troposphere, and as in the case of R1 the tropopause appeared to have no influence on these eddies. In the Tropics weak upwards transport by the eddies appeared from time to time, that shown in figure 18d being one of the better established regions.

4.3.2. *Horizontal transport.*—In figure 19 the horizontal fluxes of R2 by the meridional circulation and the eddies are presented for the first and last 10-day periods. Throughout the experiment the maximum intensity of the meridional fluxes was larger than that of the horizontal eddies, but they were more variable and do not appear to have been as effective for R2 transport as the latter. The cell structure of the mean meridional fluxes varied quite markedly with time, as a comparison of figures 19a and c indicates, with minor cells breaking into the predominantly two-cell structure. The minor

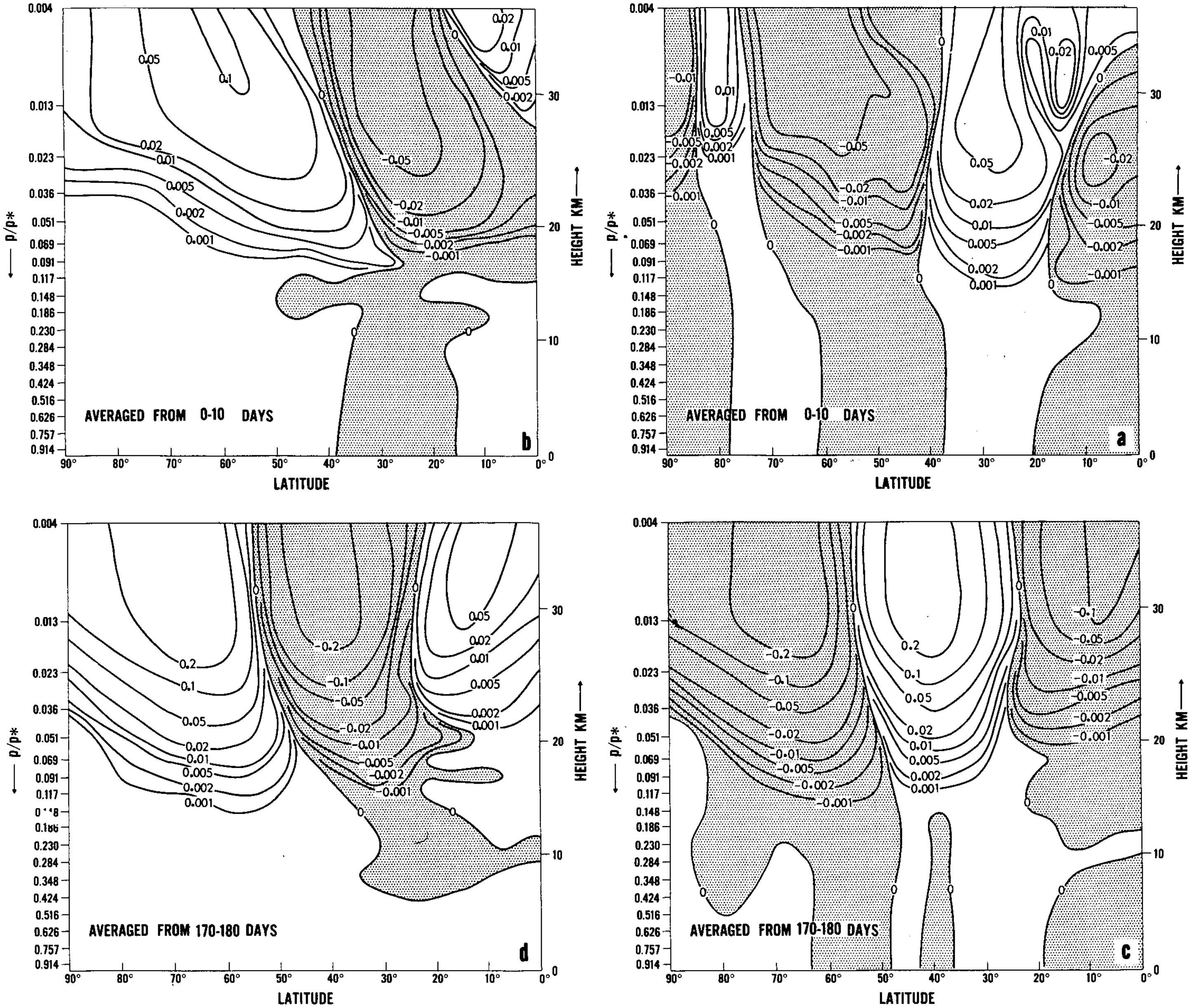

FIGURE 20.—Total rate of change of R2 by combined vertical and horizontal components. $\left(\text{Units: } \frac{\mu \text{ gm.}}{(\text{gm. sec.})} \times 10^5.\right)$ Eddies are on the left, the meridional circulation on the right. Shaded areas are regions where R2 is decreasing.

cells opposed the transports of the adjacent cells and thus tended to reduce the efficiency of the overall transport. Broadly speaking there were poleward mean meridional fluxes from the Equator to about 35°, and opposite fluxes at higher latitudes. The major fluxes were limited to the upper levels where the R2 concentrations were highest.

The horizontal transport by the eddies was relatively consistent, although the maximum intensities varied by up to a factor of four. The eddies in general opposed the mean meridional fluxes, particularly when the two-cell meridional pattern dominated. The magnitude of the equatorward eddy fluxes was noticeably less than the poleward fluxes and covered a smaller latitudinal range. The eddy fluxes were also highest at the upper levels, but were operative over a greater height range than the mean meridional fluxes and dominated the transport at the lower levels. The intensities of both fluxes fell off very fast at the lower levels, particularly in the high latitudes where the R2 concentrations were smallest. Although not shown in the figure, the maximum horizontal eddy fluxes at the lower levels were in the vicinity of the tropopause gap (cf. vertical eddies).

4.3.3. *Convergences.*—In figure 20 the convergences of R2 for the combined eddies and for the meridional circulations are depicted for the two 10-day periods. At the start the meridional convergence was slightly smaller than the eddy convergence and effectively had a three-cell pattern. Within about 30 days a two-cell pattern similar to that for the last 10 days had emerged, and thereafter three cells occurred only infrequently. At about this time the eddy and meridional rates of change were nearly in balance, being approximately equal and opposite, except at the Equator where the meridional divergence domi-

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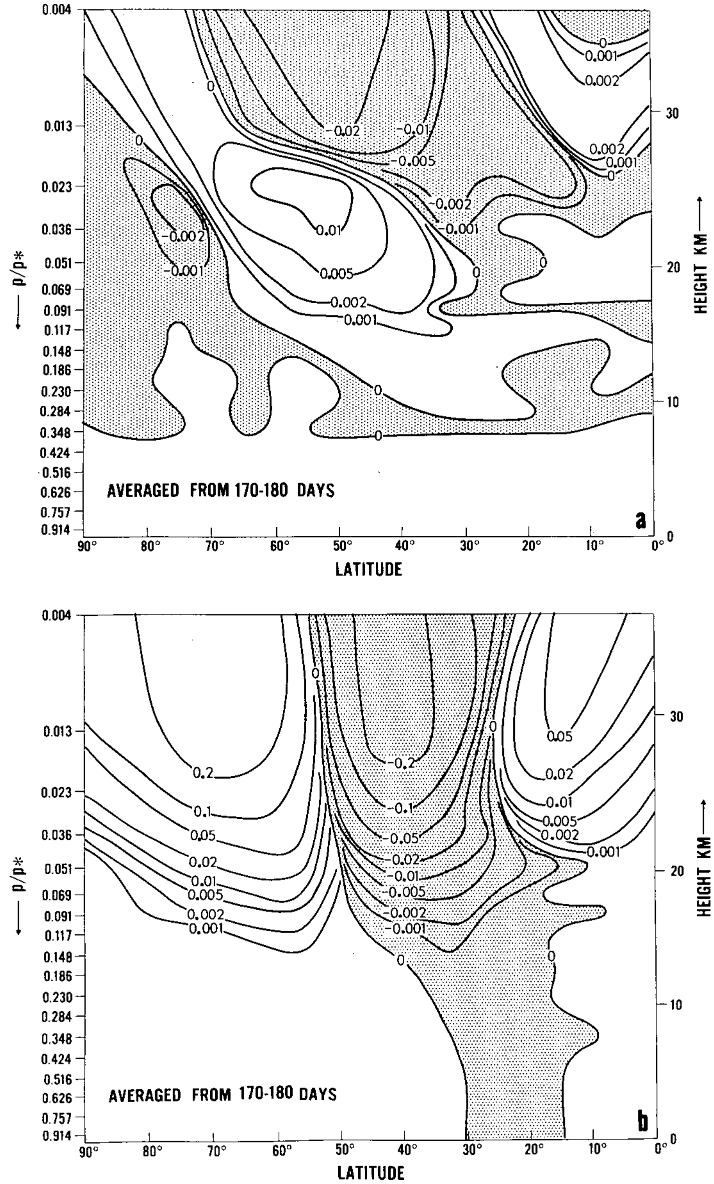

FIGURE 21.—Rate of change of R2 by the vertical and horizontal components of the large-scale eddies. $\left(\text{Units: } \frac{\mu \text{ gm.}}{(\text{gm. sec.})} \times 10^5.\right)$ Vertical eddies are on top, horizontal eddies below. Shaded areas are regions where R2 is decreasing.

nated. This state existed for the rest of the experiment, and figures 20c and d can therefore be considered as being fairly representative; because of this state of balance the net rate of change of R2 was rather small. The intensity of the mean meridional convergence increased somewhat during the course of the experiment but, following the fluxes, the maximum values were always at the upper levels. The action of the meridional circulation in transferring R2 from the equatorial and high latitudes into the subtropics is clearly shown in figure 20c, which also reveals that the convergence by the vertical meridional fluxes in the region of the tropopause gap is rather small since it is not resolved by the contouring used in the figure.

The eddy convergences did not change a great deal during the experiment, except for a gradual intensification

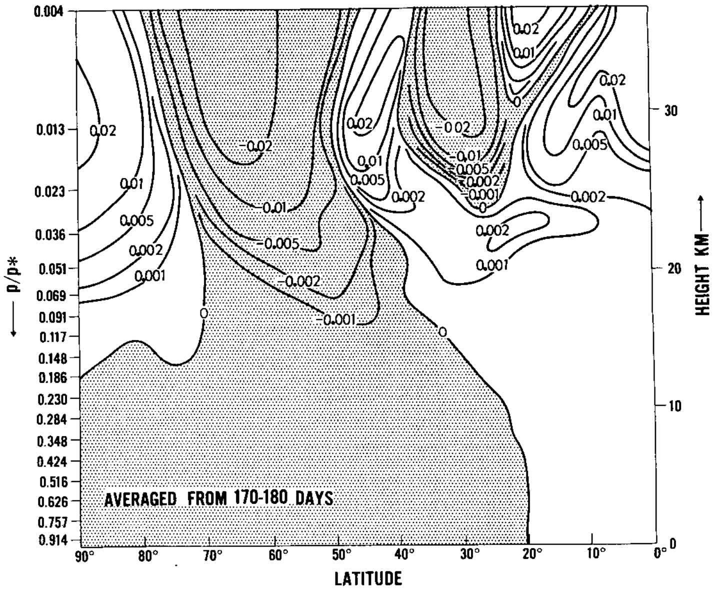

FIGURE 22.—Rate of change of R2 by subgrid scale horizontal diffusion. $\left(\text{Units: } \frac{\mu \text{ gm.}}{(\text{gm. sec.})} \times 10^5.\right)$ Shaded areas are regions where R2 is decreasing.

of the divergence area in the subtropics, notably at the lower levels. There was also a related increase in the convergence in the lower stratosphere at high latitudes. As in the case of R1, the eddies actively opposed the meridional circulations at all points and thus transported R2 from the subtropics into high latitudes, and also less effectively into the Tropics.

For completeness the total eddy convergence is broken into vertical and horizontal components in figure 21 for the last 10-day period. Throughout the horizontal eddies were about an order of magnitude greater than their vertical counterparts, and thus dominated the eddy transport as a comparison of figures 20d and 21b indicates. The horizontal convergence of R2 settled down quite fast, but because of the development of the downward eddy flux cell in the subtropics (see fig. 18) the convergence pattern of the vertical eddies evolved over a period of several weeks. The maximum rate of change of R2 by the vertical eddies was along the zero line of the horizontal eddy convergence at about 50° lat.

In figure 22 the rate of change of R2 due to the subgrid scale diffusion is given. This diffusion produced changes comparable in magnitude to those of the vertical eddies, and approximately supported the meridional divergences except in the Tropics where it supported the weaker eddy convergence. The various mechanisms transporting R2 are compared for two levels in figure 23 as a function of latitude. Unlike R1 (see fig. 11) the sign of the convergence of the major transports does not change with altitude, but the magnitude reduces drastically at the lower levels. The close balance between the meridional circulation and the eddies is apparent from the magnitude of the residual, approximately indicated by the net rate of change curve in the figure. The latter illustrates the net loss of tracer experienced in the Tropics and the resulting accumulation at higher latitudes, this accumulation being closer to the

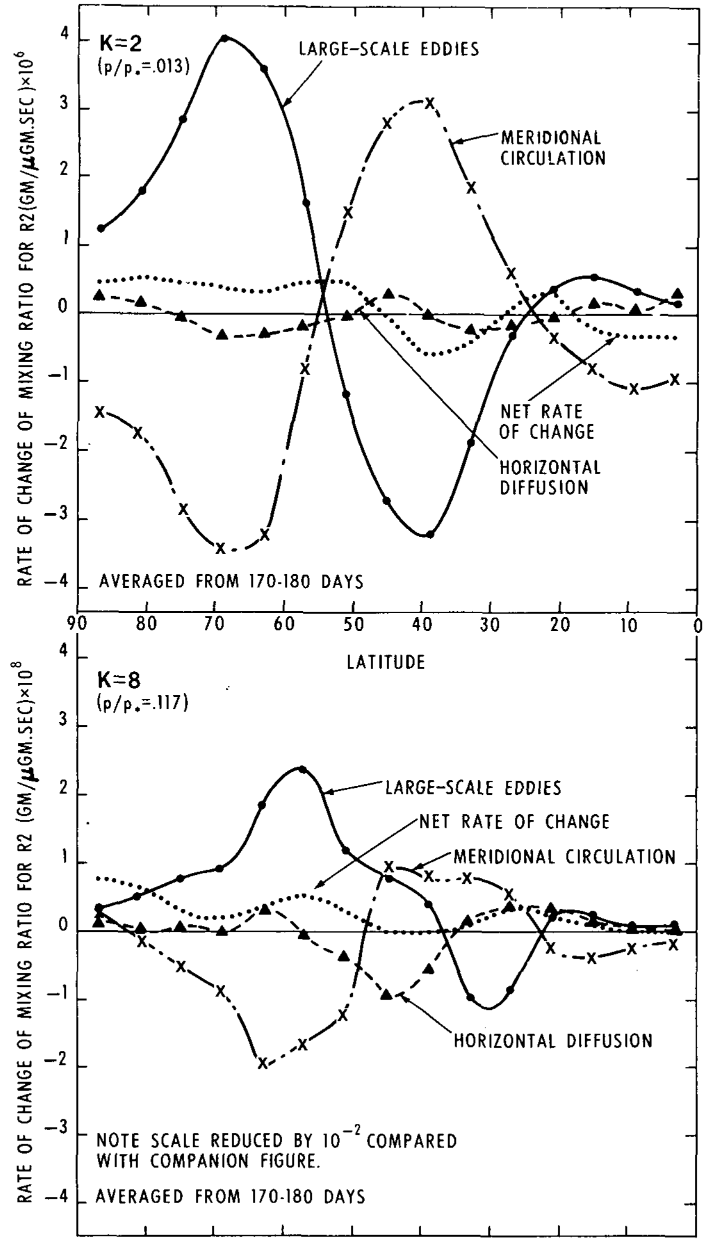

FIGURE 23.—Rate of change of R2 versus latitude for the various transport mechanisms averaged over the last 10 days of the experiment. Results are shown for levels 2 and 8.

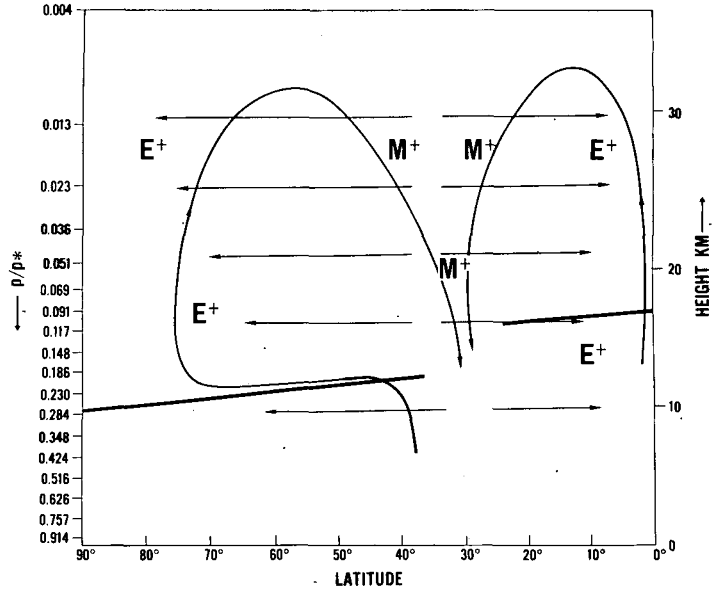

FIGURE 24.—Schematic diagram showing the combined actions of the large-scale eddies and the meridional circulation for quasi-steady state conditions for R2. E^+ and M^+ represent regions of convergence by the eddies and meridional circulation, respectively. The horizontal arrows indicate the horizontal eddies.

Equator at the lower level. The change in R2 produced by the subgrid scale horizontal diffusion is of comparable magnitude to that of the net rate of change, but this horizontal diffusion is effectively only redistributing the tracer transported by the large-scale motions, and it is the latter which are primarily responsible for the latitudinal increase of R2. A comparison of figure 23 with figure 19 of Part I shows that qualitatively the transport mechanisms produced nearly identical results for R2 and potential temperature. This happened because the mixing ratio and thermodynamic equations are very close formally, and since the R2 and potential temperature distributions are similar, particularly as regards their vertical concentration gradients, they tend to be treated in a related manner by the dynamics.

4.4. EXPLANATION OF THE R2 DIFFUSION

Based on the results and discussion given previously for R1 the diffusion of R2 might be thought explicable by essentially the same mechanism. While this is partially true, the situations are not exactly parallel and a comparison of the total eddy convergences for R1 and R2 in figures 8f and 20d reveals a very important difference. Although the figures are somewhat similar at low levels, at the upper levels each has a three-cell pattern in which the signs for R1 and R2 are reversed. That is, where there was convergence of R1 in the subtropics by the eddies previously, there is now divergence for R2. This resulted from a combination of the smaller latitudinal concentration gradient, the different vertical distribution of R2, and the much more widespread initial distribution. The latter effectively shifted the principal R2 convergence by the horizontal eddies polewards compared with R1, while it also permitted the indirect meridional cell to influence the R2 transport from the start, unlike the situation with R1. The vertical R2 distribution was also very important for the meridional circulation, as having the maximum concentrations in the top level meant that convergence by the downward branches occurred at all levels, whereas for R1 these branches had first to pass through the region of maximum concentration around levels 4 and 5 before convergence occurred. Other differences can readily be appreciated and these point to the difficulty of trying to generalize the results from any given situation. Superficially the transport of R2 is very simple to explain. Consider the convergence patterns given in figure 20c and d for the last 10-day averaged period of the experiment. These show that at all heights in the stratosphere the upward branches of the meridional circulations were removing R2 from the Tropics and high latitudes and converging it in middle latitudes and the subtropics. On the other hand the eddies were operating in the opposite direction and the two transport mechanisms nearly neutralized one another. This situation is summarized by the schematic diagram in figure 24. Comparison with the corresponding diagram for R1 in figure 12 reveals that the principal difference is the reversal of the direction of the eddy fluxes in the subtropics at most heights in the stratosphere. A balanced state was not maintained between the eddies and the meridional circulations, especially in the early part of the experiment, and the net result was that R2 was transported from the Tropics to high latitudes and accumulated there. This was possible because most of the tracer converged into the subtropics by the meridional circulations was transferred polewards by the horizontal eddies, and the resulting convergence at high latitudes was always slightly greater than the corresponding divergence produced by the rather weak indirect meridional cell. In the early stages this convergence was aided by the latitudinal concentration gradient of R2, especially in the lower stratosphere where the R2 concentrations were very small at high latitudes. Now, in the Tropics this situation was reversed as the convergence there by the horizontal eddies, which were removing R2 from the subtropics, was much smaller than the divergence by the stronger direct meridional cell. Therefore, the upward branch of this cell removed R2 from the Tropics faster than it was replaced, and transferred it to the subtropics where it combined with the downward branch of the indirect cell. Thus the subtropics acted as an intermediary in the transfer of R2 from the Equator to the high latitudes.

The final R2 distribution in the vicinity of the tropopause was very similar to that for R1, and the details as to how this was arrived at correspond closely and will therefore not be repeated here. The very uniform horizontal R2 distribution at high latitudes was primarily due to the eddy transports, but there was also a contribution from the subgrid scale diffusion which was working to smooth out local variations in concentration. The general tilt of the R2 isopleths at low latitudes can be attributed to the action of the direct meridional circulation in much the same way as for R1. The variation of the R2 concentrations in the top level during the experiment was rather confusing initially, because there were transient features present for the first 30 days which caused the concentration at the Equator to increase during this time. This was largely due to the initial R2 concentration having a maximum at about 30° lat. In addition to these transient effects truncation error associated with the vertical transports produced inaccuracies throughout the experiment, and this obscured somewhat the details of the genuine changes which occurred in this level. The overall R2 distribution was still changing quite rapidly at the end of the experiment, especially in the upper levels at the Tropics. However, the final results, as in the case of R1, leave no doubt that the net transfer mechanism of the tracer was a highly interactive combination of eddies and meridional circulations, in which no single component was supreme. This remark needs some amplification, since observational studies and current ideas indicate

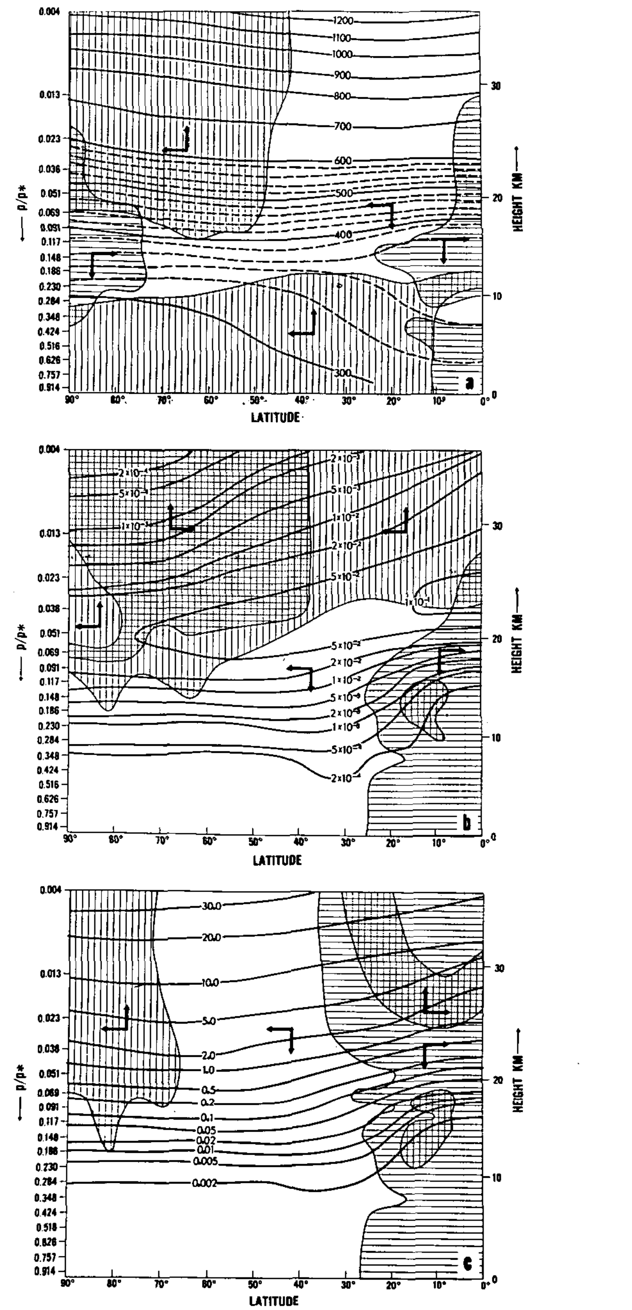

FIGURE 25.—In the upper part of the figure the directions of the horizontal and vertical eddy heat fluxes are superposed on the potential temperature field. In the middle and lower parts of the figure the directions of the eddy mixing ratio fluxes are superposed on the R1 and R2 distributions respectively. The instantaneous values of the potential temperature and mixing ratios for the last day of the experiment were used for the basic fields, while 10-day mean values for the last 10 days of the experiment were used for the fluxes. Regions of southwards and upwards fluxes are indicated by horizontal and vertical hatching respectively, regions of northwards and downwards fluxes are white. The black arrows also indicate the directions of the eddy fluxes. (Units: potential temperature, °K.; R1, gm./gm.; R2, μ gm./gm.)

that horizontal eddies are primarily responsible for the poleward transport of tracers in the lower stratosphere. These findings are not at variance with the model, but what the model does show is that the supply of tracer for the eddies is primarily provided by the meridional circulation in the subtropics. Thus the comprehensive perspective given by the model indicates that previously perhaps too narrow a viewpoint has been taken.

Although the above explanation of the R2 diffusion may seem satisfactory, closer examination of the results raises a number of questions. Why was R2 transported polewards by the horizontal eddies when for much of the experiment the horizontal gradient was equatorwards? Why wasn't the maximum concentration of R2 in the subtropics at all levels, since this is where the meridional circulations deposited it? A related question concerns the variation of the maximum concentration of R2 with latitude for the different levels of the model. There is also the question of what was the function of the vertical eddies during the experiment, and how they were related to the horizontal eddies. Since an attempt will be made to answer these questions in the following sections, only a few comments will be made here. These concern the variation of the R2 maximum with latitude and height, which can be partially explained in terms of the meridional circulations. At the start of the experiment most of the downward meridional flux was at about 27° lat., which suggests that the direct meridional cell was responsible. As R2 accumulated at high latitudes the contribution of the indirect cell increased, resulting in the latitude of maximum downward meridional flux moving to about 40° at the higher levels, but clearly remaining at about 27° for the lower levels (see fig. 18c). The vertical eddies, although somewhat weaker, also supported this latitudinal variation of the maximum fluxes with height. Thus part of this variation can be attributed to this meridional flux movement with altitude, the horizontal eddies then being responsible for the remainder of the variation. Since the horizontal eddies were strongest at the upper levels and comparatively weak lower down, this agrees with this scheme. A careful examination of figure 15 and figure 16f indicates the variation of the maximum concentration with height.

5. SOME COMMENTS ON TRACER AND HEAT FLUXES

In figure 25a, b, and c, the heat and mixing ratio eddy fluxes are superimposed on their respective concentration fields, the flow directions (but not the magnitudes) being indicated by the arrows in each region. No attempt has been made to indicate the orientation of the total flow vector of the eddies as the data available are not suitable, but in each case it should be down the total gradient. It is difficult to imagine any realistic situation in which the net transfer by the total eddies can be countergradient. Since only the total eddy vector is downgradient this means that one of its components may be countergradient, and the occurrence of such transport has been noted in the previous sections and is clearly shown in figure 25. To illustrate the point for a particular case consider the R1 eddy fluxes in the middle stratosphere at high latitudes shown in figure 25b. These indicate that while the upwards component is downgradient the horizontal component is countergradient, but the total flux should be downgradient with the R1 being transported upwards and southwards. In the region where the slope of the isopleths is greatest between 50° and 60° lat. the vertical fluxes reach a maximum, as given in figure 6f, so that even there the downgradient transfer should be maintained. Again, figure 25c shows that, although the R2 horizontal eddy flux north of the subtropics was not directed downgradient towards the Equator, the total R2 eddy flux was downgradient; hence the direction of the horizontal flux is quite acceptable. This illustrates the relationship which is required between the horizontal and vertical eddy components in order for transfers down the gradient to exist. Although this discussion answers some of the queries raised in the previous section concerning the R2 dynamics, there is still the question of why R2 was transferred away from the low concentration convergence zone in the subtropics. Some enlightenment of this problem will be provided in the next section.

In figures 4f and 16f the isopleths of R1 and R2 have been superposed on the isentropic surfaces, and it can be seen that, apart from the lower stratosphere at middle and high latitudes, the two surfaces cross one another at all points. This is of considerable importance since it implies that for mixing by eddy processes to occur down or along the concentration gradients the flow, in general, has to cross the isentropic surfaces. If the flow is steeper than the isentropes it will result in a net downgradient vertical heat transfer by the large-scale eddies, which consumes eddy kinetic energy and produces eddy available potential energy. Since the latter is dissipated radiatively the flow can be maintained only if energy is supplied from another region, and the motions are therefore considered to be forced. This cross-isentropic flow is of particular importance in the subtropics of the lower stratosphere, because the tracers pass through this region on their way to the "reservoir" at higher latitudes. In figure 25a an indication of where forcing might be expected can be obtained from the areas of downward heat flux by the vertical component of the large-scale eddies, a more quantitative estimate being given in figure 21a of Part I. The latter figure shows that the forcing is greatest in the subtropical lower stratosphere, and in agreement with this the isopleths are observed to cross the isentropes at a steep angle in this region in figures 4f and 16f. On the other hand at higher latitudes in the lower stratosphere the forcing is rather weak, and the two surfaces become almost parallel. These results are in agreement with the conclusions of Newell [30], who was the first to advance the idea that the forcing of the flow permitted the eddy transfer of tracer down the concentration gradient at a steeper angle than the isentropes.

In the upper four or five levels of the model the isentropes and isopleths of R1 and R2 also cross at a steep angle, although the forcing of the flow is very weak or nonexistent at these heights. However, in the case of the R1 distribution at these levels in low latitudes (see fig. 4f) the crossing is not of any dynamic consequence in this

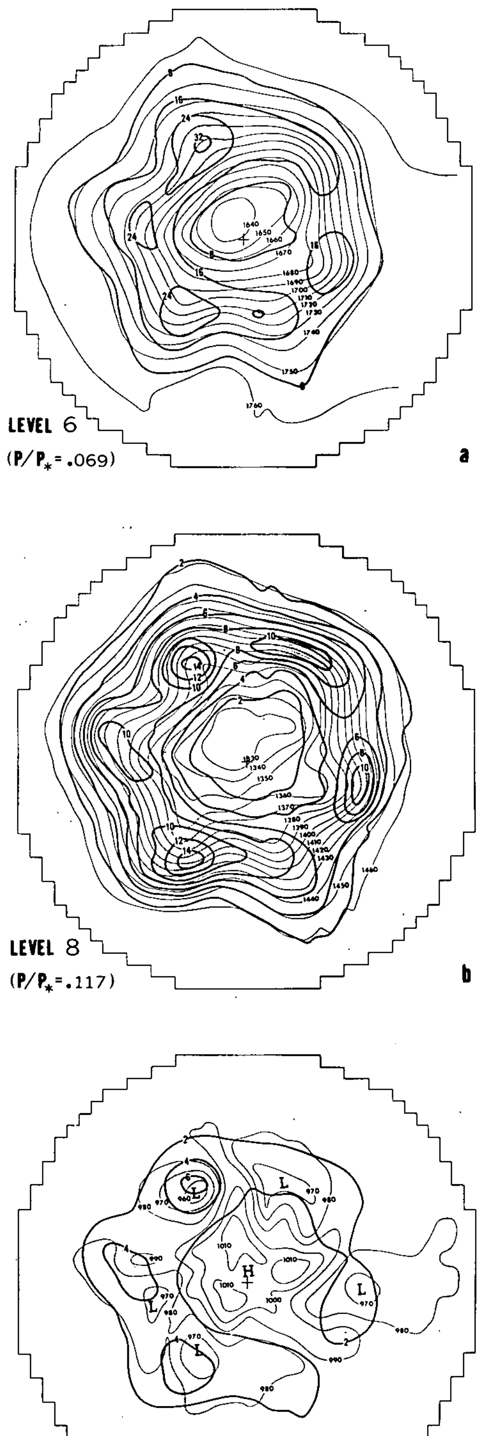

FIGURE 26.—In the upper part of the figure the R2 concentrations for level 6 are superposed on the geopotential surface for that level, while in the middle part of the figure a similar superposition is shown for R1 and level 8. The lower part of the figure shows the superposition of the R2 concentrations for level 8 on the surface pressure pattern. (Units: geopotential surfaces in meters, surface pressure in millibars, mixing ratios in gm./gm. times an arbitrary scaling factor, for each level.)

particular case. This is because the eddy motions are essentially horizontal in this region, owing to the smallness of the vertical component, and the eddy diffusion of the tracer occurs along the isentropes which are also nearly horizontal. Hence, owing to the location of the initial R1 distribution, the large-scale diffusion results in a crossing of the isentropes and isopleths in this region, without necessarily involving any cross-isentropic flow by the eddies themselves. At higher latitudes it is clear from figure 25b that the eddy motions involve crossing the isentropes, but since the flow no longer remains adiabatic this is permissible. According to figure 25 of Part I, high radiational damping exists at these levels and individual air parcels adjust their temperatures to those of the environment. This damping is presumably important for the eddy diffusion of R2 at all latitudes at these heights for the situation shown in figure 16f and 25c.

It was pointed out previously with regard to figure 23 that the thermodynamic and mixing ratio equations are very similar formally, so that some relationship between the eddy fluxes of heat and mixing ratio might be expected. However, since the eddy heat fluxes should be a more basic characteristic of the general circulation than the eddy tracer fluxes, it might be expected that the horizontal and vertical components of the tracer fluxes would have the same directions as those of the corresponding heat fluxes. A comparison of the fluxes in figure 25 reveals that while in some regions of the atmosphere this is so, there are other regions in which all three quantities have a different orientation for their total flux vector. An examination of hemispheric maps for the various quantities involved permits a qualitative assessment as to how the fluxes originate in most cases, and it happens that the different horizontal flux directions are due to both the mixing ratios and the temperatures having dissimilar horizontal distributions, which correlate differently with the wind components. The directions of the vertical fluxes of R2 and heat agree quite well in general, because of the similarity of the vertical distributions of R2 and potential temperature. The R1 vertical flux on the other hand has the opposite direction to that for the heat in the middle stratosphere at low latitudes, and this is because its vertical concentration gradient produces the reverse correlation with the vertical wind to that of potential temperature. The fluxes shown in figure 25 are therefore internally consistent, and they indicate the danger of trying to deduce flow directions for atmospheric constituents when only the heat fluxes are known.

An important feature of the present results is that in the lower stratosphere from the subtropics to about 70° lat., the flux directions are the same for all three quantities, and they indicate that tracers and heat are being transferred polewards and downwards. This is in agreement with the results of Newell [29] for ozone and heat fluxes in this region of the actual atmosphere.

6. SYNOPTIC FEATURES

It is known from the work of Dobson et al. [7] that the local total ozone amount in the atmosphere is related to the presence of individual cyclones and anticyclones, the corresponding influence of the associated upper level troughs and ridges having been studied by Normand [31]

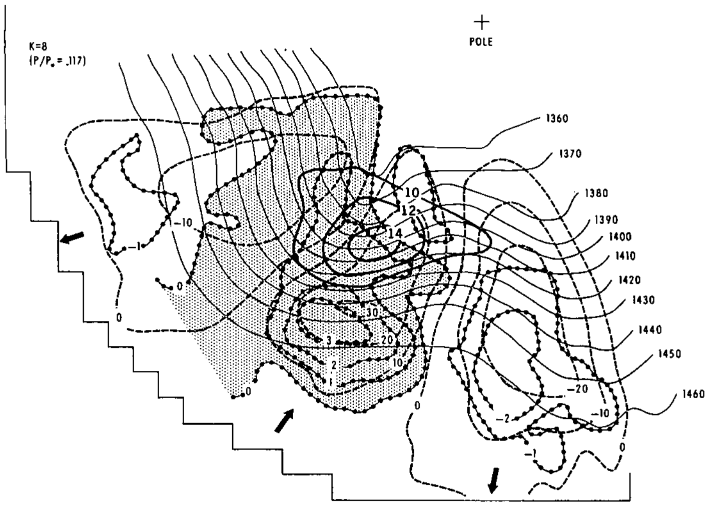

FIGURE 27.—An analysis of a trough shown in figure 26. The thin solid lines correspond to the geopotential surface, the heavy solid lines to the R1 tracer, the dashed lines to the meridional velocity, and the dotted line to the vertical velocity. The shaded region is an area of downwards motion, and the short black arrows indicate the direction of the meridional velocity at the different locations. All quantities are instantaneous values at 66 days after the tracers were initialized. (Units: geopotential surface in meters, mixing ratio in gm./gm. times an arbitrary scaling factor, meridional velocities in m./sec., and vertical velocities in cm./sec.)

and Reed [33]. It is therefore of interest to investigate if similar results are obtained in the model for R1 and R2, even though the model resolution is rather poor for such a study. The R2 concentrations are, of course, much lower than those of ozone throughout the lower stratosphere, but this should not affect the qualitative behavior of R2 so that a meaningful comparison can be made. No synoptic scale observations are available for radioactive debris in the atmosphere, but there is every reason to assume it would respond similarly to ozone and, as far as the model is concerned, there is no reason to discriminate between R1 and R2. Results for both tracers will subsequently be given and it will be assumed that they are equally applicable to either.

In figure 26a and b the local R1 and R2 concentrations for different levels are compared with their corresponding geopotential surfaces for an arbitrary time, which was chosen to be 66 days after the tracers were initiated. It is clear from these figures that the local concentration maxima are closely related to the troughs of the planetary scale wave system, these troughs being the upper level components of the surface cyclone system shown in the lower part of figure 26. The trough and isopleth patterns are more intense at the lower levels but the same basic features are obviously present at all the levels shown, and illustrate the vertical continuity of the model. A general feature of these local concentration peaks is that they tend to lie slightly ahead of the trough, and this turns out to be of considerable importance in producing the northward and downward fluxes observed in the lower stratosphere. The relationship of the tracer concentration to the surface pressure system shown in figure 26c is similar in some respects to the findings of Dobson et al. [7], although the situations are not parallel. Dobson et al. superposed the departures of the local total ozone concentrations from their monthly mean values on the isobaric contours for cyclonic conditions, and found that positive ozone departures were associated with the presence of a cyclone, and negative departures with an anticyclone. Since a monthly mean value is not realistic for the nonsteady state conditions prevailing in the model, the instantaneous R2 concentrations at level 8 are superposed on the model surface pressure pattern in figure 26c. It is the variations of the R2 concentrations we wish to compare with the surface pressure, and as these variations have been shown previously to be related throughout the lower stratosphere, this comparison should be fairly representative of conditions in this altitude range. In the case of ozone, it is well known that its day-to-day variations also result from concentration changes in the lower stratosphere; hence the R2 variations should be closely related to the ozone departures noted by Dobson et al. The results in figure 26c verify this as the local R2 maxima are clearly associated with the surface Lows, while the Highs are generally located in the areas of lower concentration. In particular, especially for the intense Low in the upper part of figure 26c, there is a tendency for the maximum concentrations to occur to the west of the cyclones as observed by Dobson et al. They found a similar tendency for the minimum concentrations associated with the anticyclones, but the model results are not adequate to verify this point. The model therefore seems to be able to reproduce quite well some of the more detailed features of the atmosphere.

A breakdown of the R1 and meteorological situation for a particular trough of level 8 shown in figure 26b was made in order to investigate what relationships exist between the various terms, and this proved to be very important for understanding how the eddy fluxes of the tracer in the stratosphere are obtained. The resulting map is given in figure 27, and shows a superposition of the R1 mixing ratio and vertical and meridional velocities on the geopotential surface in the region of the trough. This figure is rather complex and needs careful examination in order to resolve the features of interest. The figure shows that the maximum R1 concentrations do not coincide with the maximum downward velocities in the subtropics, but are to the north, indicating that vertical advection is not the prime mechanism by which the R1 concentrations are maintained in middle latitudes, in agreement with the conclusions of Reed [33] on ozone. However, it is important to note that most of the region of high R1 concentration is associated with downwards motion and also fairly large northwards motion, and the resulting correlation will give a transfer of R1 which is down the concentration gradient in the lower stratosphere. Since the troughs and ridges essentially represent the eddies of the general circulation flow pattern, this transfer of R1 will be in agreement with the results discussed previously for the eddy fluxes in this region. Obviously other regions of the trough and also the adjacent ridges will produce eddy fluxes which are not in agreement with these results, but these regions are

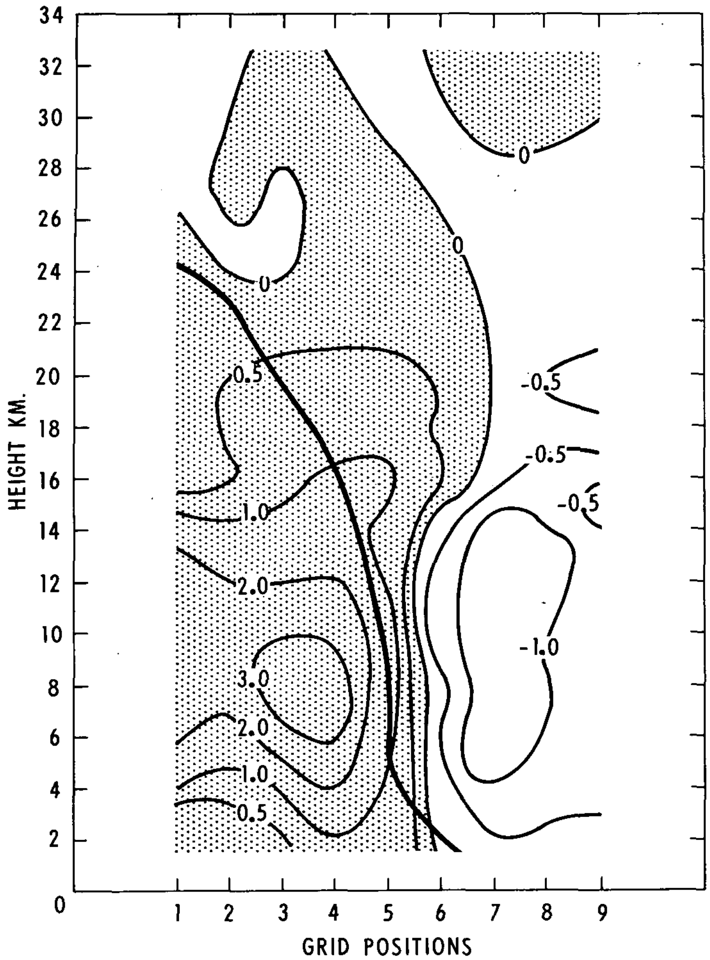

FIGURE 28.—The variation with altitude of the trough line and vertical velocities in a trough. (Units: cm./sec.) The heavy line shows the position of the trough center, the thinner lines are contours of constant vertical velocity. Downwards motions are shaded.

associated with lower R1 concentrations so that the effective transfer should be as described above. These other regions appreciably reduce the efficiency of the net downgradient transport in the lower stratosphere since, on the whole, they produce an upwards and southwards transfer. The particular trough situation analyzed here admittedly represents a rather good example of the correlations which must exist in order to produce the required fluxes, as an examination of the other troughs in figure 26 reveals, but on the average these other troughs must also behave similarly in order to produce the observed results.

In view of this it is of importance to understand why the maximum concentration of tracers occurs in conjunction with the troughs, and how the observed relationships are maintained. This situation results from the troughs being essentially regions of downwards and northwards flow in the lower stratosphere, while the adjacent ridges are regions of upwards and southwards flow, as shown in figure 27, these phenomena being a fundamental feature of the general circulation which is related to the baroclinic nature of the atmosphere. Now it has been shown previously that the supply of tracers for the lower levels is primarily maintained from the upper levels by means of the downward branch of the meridional circulation in the subtropics. However this was for zonal mean conditions, and it is apparent that when viewed on a synoptic basis the increase of tracer at a lower level will only occur in the vicinity of a trough. In the case of a ridge the vertical motion will actually produce a decrease of concentration, because of the vertical concentration gradient of the tracer. In figure 27 it can be seen that areas of maximum vertical velocity are in close proximity to those of maximum meridional velocity; hence in a trough the tracer which is transferred from above will rapidly be carried polewards. This polewards transport will produce convergence of the tracer somewhat to the north of the area of maximum meridional velocity, because of the weakening of the meridional velocity in that region. This, incidentally, accounts for the R2 tracer not accumulating in the subtropics, but being moved polewards rather than towards the low concentration regions in the Tropics. The net result of these motions is the production of an area of high local concentration on the polewards side of the trough, which means that a distinct concentration gradient exists in this region. Examination of figure 27 reveals the necessity for the tracer concentrations to be located slightly ahead of the trough in order to obtain the optimum advantage for correlations between the tracer and the velocities. It should be clearly understood that the accumulation of tracers in the lower stratospheric "reservoir" at extratropical latitudes depends crucially upon the vertical gradient of the mixing ratio of the tracers. A vertical mixing ratio that, for example, decreased with height in the stratosphere should result in a net tracer transfer upwards and equatorwards in the lower stratosphere.

Thus the displacement of the areas of maximum velocity and maximum tracer concentration in a trough, and the dieffrence in the tracer amounts between a ridge and a trough can readily be explained. Since these phenomena are related to the transport of tracers on a planetary scale, it is obvious that the synoptic situation cannot be divorced from the general circulation as far as the transport properties of the atmosphere are concerned. Although only tracers have been discussed here, it is clear that similar considerations must apply to the transport of heat and angular momentum in the atmosphere. It is obviously important to consider the atmosphere in terms of troughs and ridges, as well as eddies and mean meridional circulations, which is the current fashion. As the map in figure 27 indicates, limiting a study to zonal mean quantities only can severely restrict the insight one obtains into how the atmosphere functions.

Some of the points made in the above discussion are in agreement with the conclusions of Reed [33] concerning the explanation of the well-known ozone-weather relationship.

One of the problems connected with the polewards flux of tracers in the lower stratosphere is that the associated downwards motion should produce adiabatic heating in the troughs. Larger cooling rates than those observed in the lower stratosphere would then be required if this

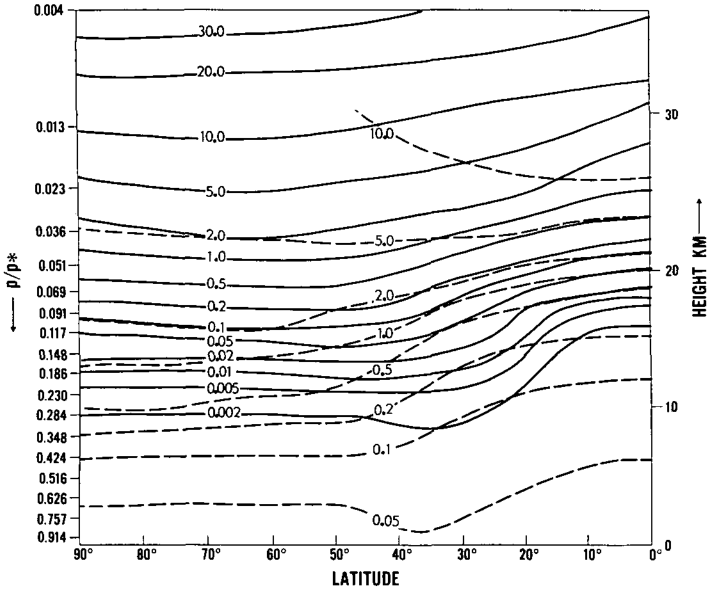

FIGURE 29.—A comparison of the zonal mean R2 distribution at the end of the experiment with the annual mean ozone concentrations obtained over the United States. (Units: μ gm./gm.) The full lines are for R2, the dashed lines for ozone.

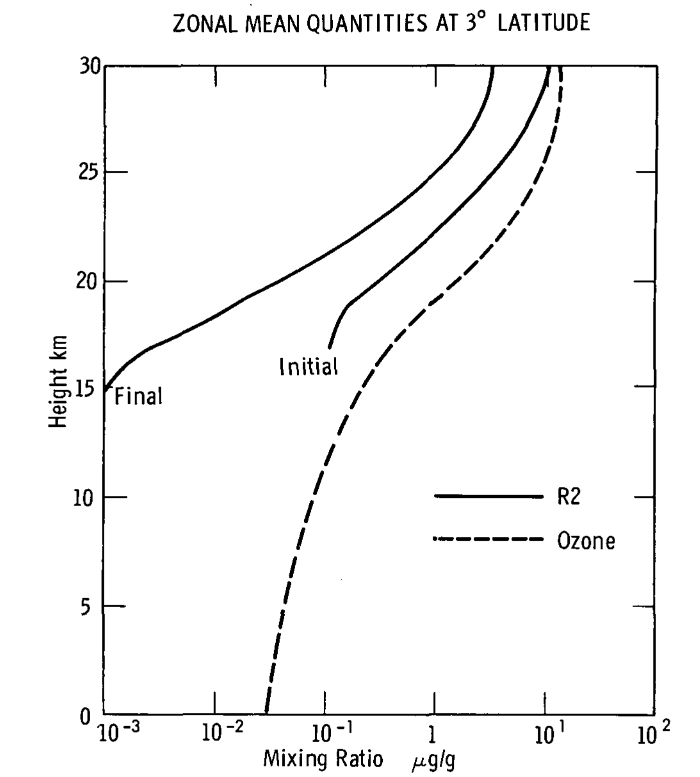

FIGURE 30.—The vertical distribution of the zonal mean R2 at 3° lat. for initial and final conditions is compared with the observed annual mean ozone distribution.

heating was entirely compensated for by radiative cooling. The strongest downwards motion, and thus the highest adiabatic warming, occurred in the subtropics according to figure 27. However, the strong polewards flow in this region produced an associated polewards flux of heat from the subtropical source, which essentially represented part of the large-scale eddy heat flux. The resulting flux divergence created a region of dynamical cooling in the subtropics, as illustrated in figure 14b of Part I, and this helped the radiation to counteract the adiabatic warming. Thus only quite small radiative coolings were obtained in the model, of the order of 0.3°C./day in the troughs and 0.1°C./day in the ridges. At higher latitudes smaller adiabatic warmings were associated with the polewards flow in the troughs, and these, together with the dynamical heating due to the heat flux convergence, appear to be required to counteract the radiative cooling. The cooling was considerably less in the ridges than the troughs, because the vertical motions in the former produced adiabatic cooling, while in the latter they produced adiabatic warming. Since the temperatures in this region are above those for radiative equilibrium (see the discussion in Part I), radiative cooling exists in both the troughs and the ridges. Although the ridges are fairly close to radiative equilibrium, this state is never obtained as the upper level trough and ridge pattern rotates, which results in a trough with its associated adiabatic warming occupying the former position of a ridge, and vice versa.

Because of the importance of the vertical motions in a trough, an analysis of a vertical cross section of the trough shown in figure 27 has been made. The section was taken perpendicular to the trough axis and was centered over the maximum of the region of high R1 concentration in figure 27. The vertical velocity field is given in figure 28, together with an indication of the variation of the trough axis with height. In the lower troposphere the tilt of the trough with height is in the direction required by baroclinic theory, and the angle of the tilt is much larger than in the upper troposphere. In the lower stratosphere the trough line again tilts quite steeply, and for the particular synoptic situation analyzed the tilt increased rapidly with height. The interesting feature of the analysis is that in the troposphere the trough line approximately divides the regions of upwards and downwards motions, with the air ascending ahead of the trough, while in the lower stratosphere there is a general descent over the center of the trough. The vertical velocities given in figure 28 are fairly large, being considerably greater than any associated with the mean meridional circulation.

An attempt was also made to relate the R2 and temperature profiles in the region of the tropopause gap, as was done by Breiland [2] for ozone, to determine whether examples of penetration of polar or equatorial air into this region could be found. However no perturbations such as those revealed by ozonesonde measurements were established, and this is thought to reflect on the low R2 concentrations and poor model resolution. Again, because of the latter, only very weak fronts are likely to be formed in the model, so that a study of tracer concentrations and their ability to reveal tropospheric-stratospheric interchange in the frontal zones, such as was studied by Berggren [1] for ozone, also could not be made.

7. COMPARISON WITH OBSERVATION

Unfortunately there is relatively little information available which can be used to check the results of the model against the atmosphere, and to evaluate it against the two-dimensional models which have been developed. In the case of R2, although considered to be an "ozone-like" tracer, in reality it represents a hypothetical tracer for which the atmosphere has no actual counterpart. This is because R2 was defined initially as a photochemical ozone distribution, but no photochemistry was included in the model. The model could therefore only redistribute the initial concentrations, and since photochemical ozone amounts are always much less than observed ozone amounts except at the Equator, quantitatively the R2 values could never approach those of ozone in the atmosphere. In spite of this it is of interest to compare the R2 and ozone distributions, and this has been done in figure 29. The R2 values are instantaneous values at 180 days after initiation, while the ozone values are annual mean mixing ratios, derived from the results of Hering and Borden [11] for the ozonesonde network over North America. Similarities are apparent in the two distributions in the lower stratosphere and troposphere, in particular the uniform mixing ratios for a given height in the high latitudes, the downward bulge of the isopleths in the subtropics, and the upward rise in the Tropics. Clearly the biggest discrepancy is in the value of the mixing ratios for adjacent R2 and O_3 isopleths. However, if the R2 isopleths were brought down into the vicinity of the O_3 isopleths of the same value then the overall agreement would be improved. Obviously the R2 concentrations throughout the lower stratosphere need to be increased considerably, but what is of greater interest is that the initial photochemical distribution for R2 in figure 16a has been transformed by the dynamics so that it now has similar qualitative features to those for O_3 in the actual atmosphere. In the Tropics the R2 isopleths rise faster than those of O_3, but this is because the upward branch of the meridional circulation removed R2 thus forcing the isopleths up. In the atmosphere presumably photochemistry replaces the O_3 removed in this way, thereby minimizing the influence of the meridional circulation on the O_3 concentrations in this region. At the upper levels the agreement is poor at all latitudes, but this can be mainly attributed to the initial R2 distribution having too large mixing ratios there. It is apparent from figure 29 that the vertical profiles of R2 and O_3 are quite different at all latitudes, and an indication of the extent of this difference in the Tropics is given in figure 30. Also shown in this figure is the initial R2 distribution, which reveals how much the R2 concentrations in this region have been reduced by the dynamics in the model. If the R2 concentrations in the Tropics had been continuously regenerated the concentrations at higher latitudes would have been expected to increase at a faster rate, and better agreement with O_3 would have resulted.

The reversal of the initial R2 latitudinal concentration gradient shown in figure 16, which occurred during the course of the experiment, although incorrect at the upper levels, is in qualitative agreement with the O_3 distribution in the lower stratosphere. This in itself is gratifying to some extent since the influence of the photochemistry is such that it will affect the upper levels where the disagreement occurs, and leave the lower levels as they are.

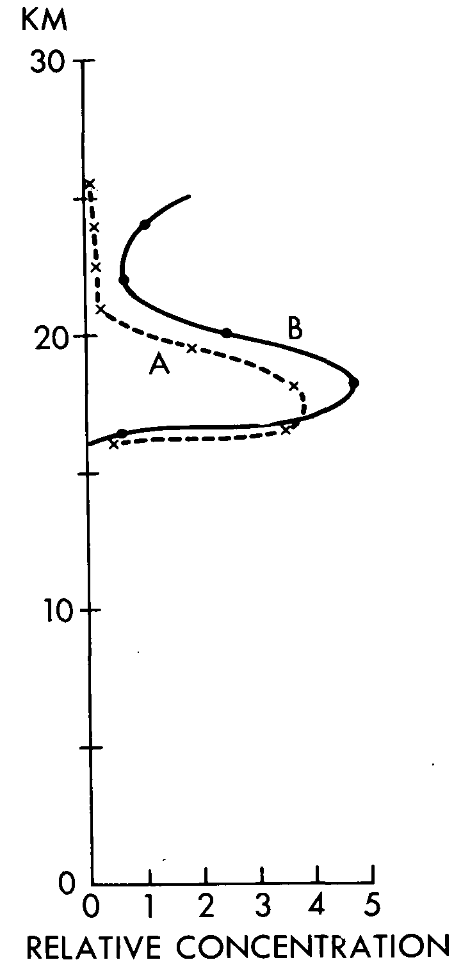

FIGURE 31.—Initial composite W^{185} mixing ratio as a function of latitude (in relative units decayed to Aug. 15, 1958). Curve A is based on a uniform W^{185} mixing ratio with height between reported nuclear cloud base and top for each event. Curve B is based on a uniform W^{185} mixing ratio within a cloud shaped like an inverted cone whose apex and base are at the reported nuclear cloud bottom and top respectively (after Machta [23]).

The reversed latitudinal gradient of R2 obtained in the experiment will, of course, also produce a variation of the total R2 amount with latitude similar to that observed for O_3. The model therefore undoubtedly had the correct tendencies as far as the development of the R2 distribution was concerned.

However, of more importance than this qualitative agreement of R2 and O_3 is the question of whether the same transfer mechanisms are responsible. Since most of the observational information is on horizontal fluxes by transient eddies, no data at all being available on the vertical transport mechanisms, this question cannot be resolved one way or another. Newell [29] has presented O_3 fluxes for the Northern Hemisphere based on total O_3 amounts, since individual vertical profiles were not available. Apart from some anomalous results for Japan, which are attributed to the presence of the jet stream, he found fairly persistent poleward transient eddy fluxes for all latitudes at 50 and 100 mb. throughout most of the year. The data sample available was very restricted for the latitude range 0–30°N., and the combined results gave a net northwards flux which was about one-third of that for the latitudes 30–60°N. This implies that around 30° lat.

an additional source of tracer must exist, which presumably is the downwards flux by both the eddies and the meridional circulation that the model shows to occur in this region. More statistically significant observations at a number of levels are required at low latitudes to confirm the magnitudes of the eddy fluxes there relative to those at higher latitudes. In the case of the model polewards eddy fluxes were obtained at all heights, except south of about 25° lat. where weak equatorwards fluxes were predicted. These equatorwards fluxes resulted from the omission of photochemistry from the model, as under these circumstances this was the only way in which the continual removal of R2 from the Tropics by the meridional circulation could be replaced. Finally the scant observational data on mean meridional fluxes agreed with the model in giving equatorwards transport over middle latitudes; however, in view of the difficulty of such observations this result cannot be considered to be established.

More accurate transient eddy fluxes for O_3 have recently been published by Hering [12] for midlatitudes based on the North American ozonesonde network. These show that there is a northward flux at all seasons at these latitudes, but that it is virtually confined to between the tropopause and 18-km. altitude. The model gave northward fluxes at all levels for middle latitudes with the maximum in level 2, and is therefore quite different from the atmosphere in this respect. The inclusion of photochemistry would be expected to modify the model results somewhat, as a different latitudinal concentration gradient would have been attained at the upper levels, and this might have produced better agreement.

As mentioned in the previous section, the model also shows good agreement with the atmosphere as far as synoptic features are concerned.

R1 is also difficult to compare with observation since it represents a very idealized tracer as far as its initiation was concerned. In addition only rather limited measurements of atmospheric radioactivity have been made because of the problems involved in such observations. The nuclear tracer most similar to R1 is radioactive tungsten-185, W^{185}, which was introduced into the stratosphere at 11°N. during the May–July 1958 U.S. nuclear test series (Martell [28]). No information is given as to the dates of the nuclear tests which introduced the tungsten thus making it difficult to define zero time for comparison with R1. The supply of tungsten to the stratosphere occurred in the form of a series of point releases, which would make its initiation entirely different to that for R1. The vertical distribution of injected W^{185} is estimated by Machta [23] and is shown in figure 31. He estimated the initial distribution using the data of nuclear cloud top and nuclear cloud base by assuming two idealized shapes of cloud, i.e., box shape (Curve A) and cone shape (Curve B). He believes that these two profiles should cover the plausible range of possibilities of initial distribution. His results show that it was confined to a relatively narrow height range in the stratosphere. The height of the maximum was at about 17.5 km., which is 3 km. lower than that of the initial R1 distribution. Because of the limited data, and the early date at which the observed distribution in figure 31 was obtained, the theoretical distribution in this figure may be a more reasonable guide to the actual W^{185} profile in the atmosphere. Apart from precipitation measurements at the surface, most of the measurements of W^{185} consist of aircraft observations from about 70°N. to 50°S. between 15–20 km. along a line of approximately constant longitude over North and South America. The data from these intercepts are usually averaged over 2-mo. periods and presented as latitude-height cross sections (see Stebbins [36]). Information over a much more extensive network is required in order to obtain a latitude-height distribution which can be considered to be a realistic zonal mean. Since no zonal measurements are available there is no information on zonal variability, but in view of the results given by the model and the known ozone variation with longitude (London [21]), such measurements would have been of great interest.

In figure 32 the observed W^{185} latitude-height distribution analyzed by List and Telegadas [20] is presented for the 3-mo. period from December 1958 to February 1959. This time period should approximately correspond to conditions in the model at about 200 days after the tracer initiation, and figure 32 should therefore be comparable with figure 4f. The units used for the W^{185} and R1 concentrations bear no relation, hence only relative values for each tracer can be compared. Several common features are apparent between the two distributions, such as the downwards and polewards development into the lower stratosphere from the low latitude source, and the rise of the isopleths and their crowding in the vicinity of the tropical tropopause. The extension of the isopleths of both tracers into the troposphere in the region of the tropopause gap should also be noted. The secondary maximum that occurs in the W^{185} concentration at about 35° in the Southern Hemisphere may correspond to the formation of a perturbation or promontory as described previously for R1 and shown in figure 1. Such individual features tend to disappear on zonal averaging, and become replaced by the general protrusion shown by the 0.1 R1 isopleth in figure 4. The biggest discrepancy is in the height of the maximum concentrations of R1 and W^{185}, and this can be attributed to having initiated R1 at a higher level.

If one accepts the theoretical distribution of the W^{185} shown in figure 31, then the maximum concentration appears to have risen from about 17.5 km. to 20 km. or greater as shown in figure 32. This would correspond to the maximum moving upward by at least 2.5 km. or greater in the 200 days between July 1, 1958, and Jan. 15, 1959. The R1 maximum rose about 3.5 km. in the 180 days of the experiment and, although this agreement with observation is very satisfactory, it must be viewed in relation to the limited observations and also the difficulty of defining a representative zero time and initial distribution for the W^{185}. The proof of the existence of such upwards motion in the Tropics would help to verify the existence of the tropical meridional cell in the lower stratosphere. The situation is complicated by the occurrence of gravitational settling in the atmosphere, a feature not

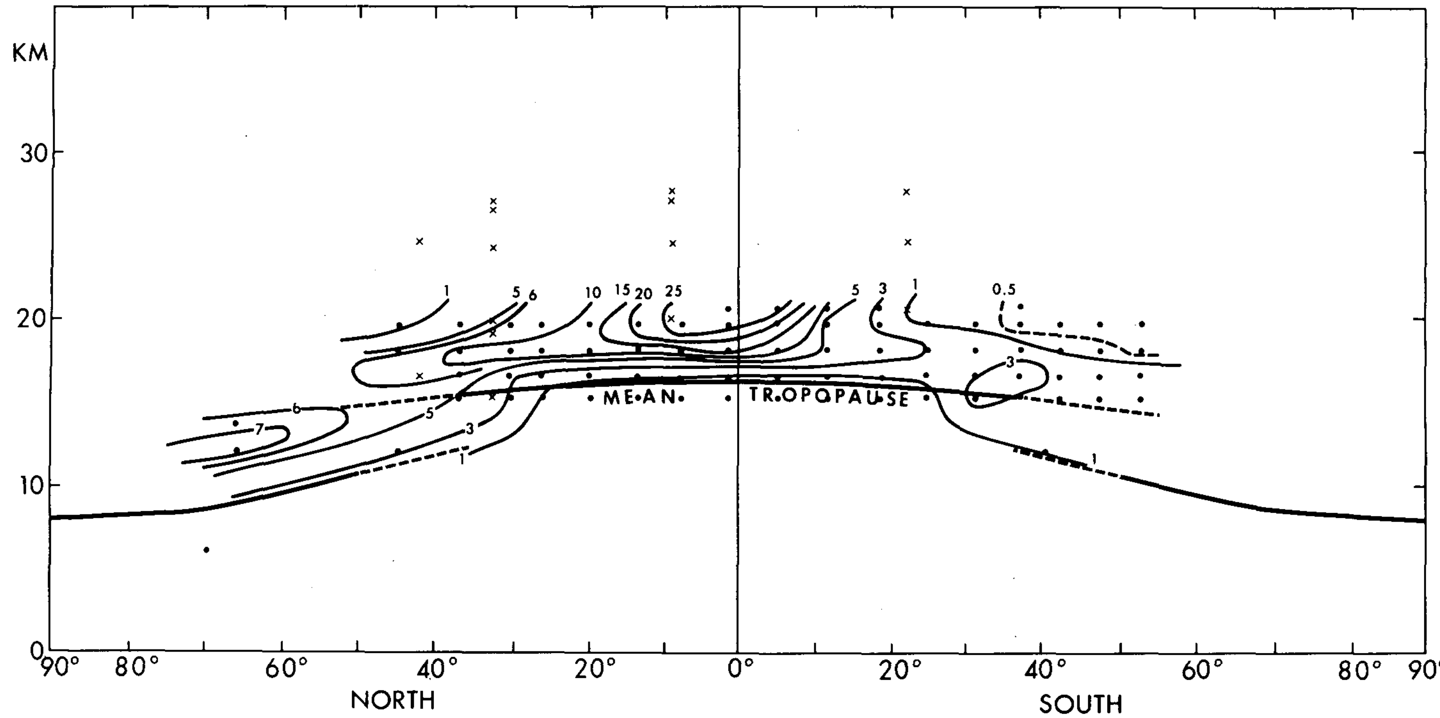

FIGURE 32.—The measured W^{185} distribution in the atmosphere based on observations made over North and South America. The dots indicate aircraft observations (mean for the 3-mo. period), while the crosses indicate balloon observations (instantaneous). (After List and Telegadas [20].) (Units: disintegrations per minute per standard cubic foot of air decay corrected to Aug. 15, 1958.)

incorporated in the model. Limited data exist on the size distribution of radioactive aerosols collected in the stratosphere. Drevinsky and Pecci [9] found that the bulk of the radioactivity from nuclear tests resides on particles in the size range of 0.02 to 0.15-micron radius. Particles of size 0.1 micron with a density of 2.5 gm./cm.^{3} would take about 6 mo. to fall 1 km. at 20 km. (Junge et al. [16]). It would therefore appear that particle settling is not of overriding importance in altering the radioactive distribution at and below about 20 km. on the time scale being considered here. The observation that the maximum concentration of W^{185} remained near the Equator is also not inconsistent with the existence of a meridional circulation in the Tropics. Owing to seasonal variations, the tropical meridional cells of each hemisphere would be expected to progress backwards and forwards across the Equator as they followed the heat Equator. This would tend to produce alternate north and south velocities at the Equator by the meridional circulations, which would result in little or no net meridional transport for a tracer in the vicinity of the Equator. The spread of the radioactive debris into the Southern Hemisphere can also be explained by assuming that the initial diffusion was primarily by the horizontal eddies, as in the case of R1, which transferred the debris down the high concentration gradient existing at this time and thus across the Equator. Once the debris was in the region of influence of the direct cell of the Southern Hemisphere's meridional circulation, it would be transferred polewards, and then downwards in the subtropics from where the horizontal eddies would continue the polewards transport.

Some evidence for the existence of meridional cells in the stratosphere has been provided by analysis of other radioactive tracers. List et al. [19] have provided some interesting information on the transport of nuclear tracers released at extremely high altitudes. In the case of the rhodium-102 tracer, they report that in the winter months it descended at about 1.5 km./mo. in the lower stratosphere north of 35°N. This amounts to a mean velocity of about 0.07 cm./sec., which should be compared with the 100-day mean value of 0.05 cm./sec. given by the model in this region. They suggest that some of their observations are best explained in terms of mean meridional transport, and presumably a model combining eddy and mean meridional fluxes similar to that given here could be hypothesized to explain the diffusion of tracers in the upper atmosphere. In the case of the strontium-90 tracer, Stebbins [36] and Feely et al. [10] have presented latitude-height distributions for this tracer for various years, although the measurements available for the latter years are hardly sufficient to permit any model to be evaluated. Krey [17] states that the seasonal variation of strontium-90 requires transport by mean meridional motions to be considered in addition to that by eddies. In particular he estimates that the downwards vertical velocity in winter at 31°N. in the lower stratosphere is 0.05 cm./sec., in agreement with the model. He also suggests that in summer there should be an upwards velocity in this region of about the same magnitude.

The emphasis here on the transport of tracers by meridional circulations in the stratosphere should not be interpreted as meaning that they are more important than eddies in the large-scale diffusion process. The model undoubtedly supports the general consensus that tracers are transported primarily by quasi-horizontal eddies in the lower stratosphere, and although there are no flux measurements available for radioactive debris in the atmosphere, there is no question but that they would be similar to those for ozone. What the model does reveal, however, is that the transport processes in the atmosphere are likely to consist of an interrelated combination of eddies and meridional circulations, a conclusion also reached by Mahlman [25], based on a study of the transport of radioactivity in the actual stratosphere. Since the general dynamical features of the model correspond to a winter situation, the transport mechanisms described here can only be considered to be valid for the autumn-winter period of the year. A considerable part of the stratospheric transport seems to occur during this time; hence the model should account for the major features of the observed tracer distributions.

Finally mention should be made of a parameter commonly used in studying radioactive tracers in the atmosphere. This is the half-residence time, which is the time for the total amount of debris to decrease by half. This time cannot be realistically estimated in the present model because of the fictitious source which existed owing to truncation error. This source acted in opposition to the precipitation, and in the early stages was the larger; thus the rate at which the model was actually losing tracer was somewhat obscured. Relatively little R1 was, however, removed from the model stratosphere, as an examination of the curves in figure 3 reveals. Despite the distortion caused by the truncation error it would seem that the model stratosphere had a noticeably longer half residence time than the actual stratosphere, and this appears to be one of the biggest discrepancies between the model and observation. This is probably partially due to the lack of seasonal variation in the model. Also in the atmosphere dust and water vapor must exert a marked influence on any radioactive debris with which they come in contact, and this may affect the fallout rate. Thus it is thought that these omissions rather than a basic flaw in the model are responsible for the discrepancy.

8. COMPARISON WITH OTHER MODELS

No previous studies have been made of tracer diffusion in the atmosphere using a complete general circulation model so a direct comparison is not possible. However, it is of interest to compare the present model with two-dimensional models which have been developed or proposed, even though this is hardly a realistic procedure.

Currently the most actively studied two-dimensional models are eddy diffusion models, in which the atmospheric transports are defined in terms of the gradient of the tracer mixing ratio and somewhat arbitrary eddy diffusion coefficients. Recently such models have been presented by Davidson et al. [6], Machta [24], and Reed and German [34], all of which were used to study the diffusion of radioactive debris in the atmosphere. Despite the exclusion of transport by the meridional circulation these models are able to account for the gross features of the spread of debris, provided the correct assumptions are made concerning the distribution of eddy diffusion coefficients. The effectiveness of these models is illustrated by the work of Davidson et al., who predicted the principal features of the diffusion of W^{185} fairly satisfactorily for a period of about 1 yr. Because of the speed and simplicity of this type of model, it is ideally suited to a task of this nature, even though a considerable amount of "curve fitting" was required in order to obtain acceptable results. These models cannot be considered to really reproduce the behavior of the atmosphere as regards the transport of tracers, especially as there appears to be considerable evidence pointing to the importance of transport by the meridional circulation. In addition, since the eddy diffusion coefficients are specified, no information is provided by such a study concerning why the atmosphere transports tracers in any particular fashion, so no insight into the behavior of the atmosphere is obtained.

A number of models (Brewer [2], Dobson [8], Libby and Palmer [18], and Machta [22]) have been advanced to explain the transport of tracers in terms of the meridional circulation. These models are hypothetical since they only represent possible transport processes, and no numerical calculations have been made with them. Only a qualitative comparison is therefore possible with the present model, and this reveals that none of these models has the two-cell stratospheric circulation obtained here, although they all have upward motion in the Tropics. A meridional circulation only model also has difficulty in accounting for the angular momentum balance of the atmosphere, and this restricts the validity of any deductions made from such a model. The basic failing of these models is that they cannot really be used, since they provide no means of calculating the magnitude of the meridional velocities required for their proposed circulation patterns. It is of course possible by trial and error to choose velocities which produce tracer distributions in approximate agreement with observation, but this procedure suffers from the criticism leveled against the eddy diffusion models.

Finally, mention should be made of the more complex models that have been developed by Prabhakara [32] and Byron-Scott [4] for deriving the ozone distribution in the stratosphere allowing for both photochemistry and transport processes. Prabhakara combined a two-dimensional eddy diffusion model with subjectively chosen meridional velocities and obtained quite satisfactory ozone amounts, although these largely resulted from the choice of the parameters in his model. The most complex calculations carried out so far for tracer studies are those by Byron-Scott [4], who incorporated ozone photochemistry in his model and investigated the joint photochemical-radiative-dynamical interaction. His model was limited to the stratosphere and was based on the three-dimensional NWP grid, the equations being scaled and the stratosphere forced from the lower boundary of the model. There is very little observational evidence available concerning this forcing term, because of the difficulty of obtaining accurate estimates for vertical motion, and it is therefore difficult to judge how much this influenced his results. He was mainly concerned, in any case, with problems related to stratospheric warmings rather than to explain how tracers are diffused in the atmosphere, but he noted that planetary waves were important for the transport of heat and ozone.

In conclusion it might be noted that the general circulation study reported here indicates that both types of two-dimensional models discussed above seem to contain part of the mechanism by which the atmosphere transports tracers. Although it is a very desirable aim, there appears to be no way in which a simple joint model can be made without arbitrarily defining the parameters involved.

9. CONCLUDING REMARKS

The study of the diffusion of tracers in the atmosphere using a general circulation model has revealed several important features, which it is thought may be representative of winter conditions in the actual atmosphere. Foremost of these is the close relationship and interdependence of the eddy and meridional circulation transports, which have been shown to be approximately equal in magnitude and opposite in action for quasi-steady state conditions. A direct meridional cell appears to be of fundamental importance in the Tropics, and it supplies a natural explanation for the observed crowding of tracer isopleths just above the tropical tropopause and their gradual dispersion at higher altitudes. In addition, the development of the tracer isopleths in the subtropics is attributed partially to the downward branch of the meridional circulation together with a comparable contribution from the vertical eddies. A weaker, indirect cell was shown to exist at higher latitudes in the stratosphere, and it appears to be of importance for the transport of tracers from the lower to the higher stratosphere at these latitudes, and if present in the actual atmosphere should exert some influence on the removal of radioactive debris from the polar reservoir. In agreement with observation, quasi-horizontal eddies were found to be of greatest importance for the transport of tracers in the lower stratosphere, where they were directed down the concentration gradient. The eddies dominated the diffusion where high concentration gradients existed, but in more general situations it was not always clear in which direction the eddy transfer would take place, making it difficult to generalize the results. A particularly noteworthy feature of the model is the information provided by the synoptic analysis concerning how the general circulation transports arise. The fundamental relationship of these transports to the upper level trough and ridge system and the surface pressure pattern has also been illustrated. In view of the agreement of the model synoptic relationships with those of the atmosphere, there is good reason to hope that the basic transport model presented here is sound.

With regard to the question of tropospheric-stratospheric exchange, it was found that this occurred principally in the region of the tropopause gap in the subtropics, with the downward branch of the direct cell and the vertical eddies being of about equal importance. (The model does not have a polar front break because its resolution is too coarse, so that the transport which is known to occur in this region in the atmosphere was missing in this study.) The supply of tracer for middle and high latitudes in the troposphere was mainly by horizontal eddies transporting from the subtropical penetration, vertical mixing across the tropopause being of minor importance.

The large-scale dynamics in the model reproduced qualitative features of the ozone distribution in the actual atmosphere, such as the discontinuity in the vertical concentration profile which occurs at the tropopause. The downward bulge of the O_3 distribution in the lower stratosphere north of the subtropics was obtained, and since this is the reservoir region for O_3 it helps to maintain the observed latitudinal increase in the total O_3 amount.

Observationally better data are required to check out the model. In the case of ozone, flux measurements for a number of latitudes and heights are required for different seasons, and a detailed analysis of a trough and ridge pattern using ozonesonde observations would be of use in verifying the picture given by the model. For radioactive tracers a definitive result concerning the influence of upward motion in the Tropics is of vital importance, as this will clarify considerably the role of the meridional circulation in the large-scale transport of tracers.

The model presented here is not considered to be representative of the summer stratosphere since it cannot explain the easterly winds which exist at this time, and the transport of tracers is presumably due to a different eddy and meridional circulation system. The model should however be fairly satisfactory for winter conditions and also the spring, when the additional supply of energy from the troposphere probably destroys the quasi-balance of the eddies and the meridional circulation in favor of the eddies, thus permitting the well-known accumulation of tracers in the lower stratosphere to occur at this time.

Further experiments are desirable to check out the model, such as the release of tracers at different altitudes and latitudes and also, of primary importance, a seasonal march. Experiments with a model of sufficient horizontal resolution to resolve fronts in the atmosphere would be of value in studying tropospheric-stratospheric interchange. Finally, a test involving a model without a wall at the Equator, and including the influence of moist processes in the general circulation, would determine whether the present model has overestimated the influence of the meridional circulation in the Tropics.

At the present time a further experiment is underway in which the model has been started from a photochemical ozone distribution, but this time the photochemistry of the ozone has been included. It is hoped that this will result in realistic ozone concentrations, thus permitting the model ultimately to calculate its own ozone distribution for use in the radiative transfer calculations.

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