Interhemispheric mass exchange from meteorological and trace substance observations

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ABSTRACT

Two mechanisms contribute to interhemispheric mass exchange: large scale eddy transfer and mean meridional circulations. Estimates of both are presented based on data from a recently completed study of the general circulation of the tropics. The long term interhemispheric exchange time is found to be about 320 days. It is suggested that when eddy processes reinforce the transfer by mean meridional circulations in the upper troposphere in late spring and summer much faster exchange rates may occur which may be responsible for an apparent anomaly in the seasonal variation of surface ozone in the southern hemisphere. Finally the interhemispheric mass flux is compared with a crude estimate of the tropospheric-stratospheric mass flux based on the same tropical data.

Introduction

The rate at which mass is transferred across the equator and the mixing ratio gradient are the governing factors in the space-time distribution in one hemisphere of a trace substance introduced into the other hemisphere. Examples of such substances are CO_2, introduced primarily into the northern hemisphere by both natural and man-made sources, radioactive fission products until recently mainly originating in the northern hemisphere, volcanic eruption debris, in recent years introduced mainly into the southern hemisphere, and numerous gases and aerosols originating primarily in the industrial regions of the northern hemisphere. Estimates of the exchange rate have hitherto relied upon measurements of a trace substance in both hemispheres (e.g. Junge, 1962*a*; Junge & Czeplak, 1968) but with the recent extension of large-scale general circulation studies across the equator to 30° S it is possible to base an estimate solely upon meteorological observations for comparison. In this note data is presented on interhemispheric mass transfer by mean meridional motions in the tropical troposphere and on possible transfer by eddy processes up to heights of 30 km.

Transfer by mean motions

The procedures used to obtain patterns of mean meridional motions and the associated mass transfer in the direct Hadley cells have been reported elsewhere (Kidson, Vincent & Newell, 1969). Herein streamlines from these flows are reproduced for four three-month averages (Fig. 1); they are based on approximately eight years of data from 300 stations in the latitude belt 40° N–30° S. Note that in December–February and June–August the cells span the equator and are therefore very effective in interhemispheric mass exchange. In March–May the air circulates in two cells essentially within a given hemisphere whereas in September–November the pattern is somewhat in between that for March–May and that for June–August with a transfer from the northern hemisphere to the southern hemisphere across the equator in the upper troposphere. The total mass crossing the equator is as follows:

Dec.–Feb.: 1.15×10^{14} g sec^{-1}
Mar.–May: 0
June–Aug.: 1.40×10^{14} g sec^{-1}
Sept.–Nov.: 0.75×10^{14} g sec^{-1}.

The long term average flux is thus
$(1.15 + 0 + 1.40 + 0.75)/4 = 0.83 \times 10^{14}$ g sec^{-1}.

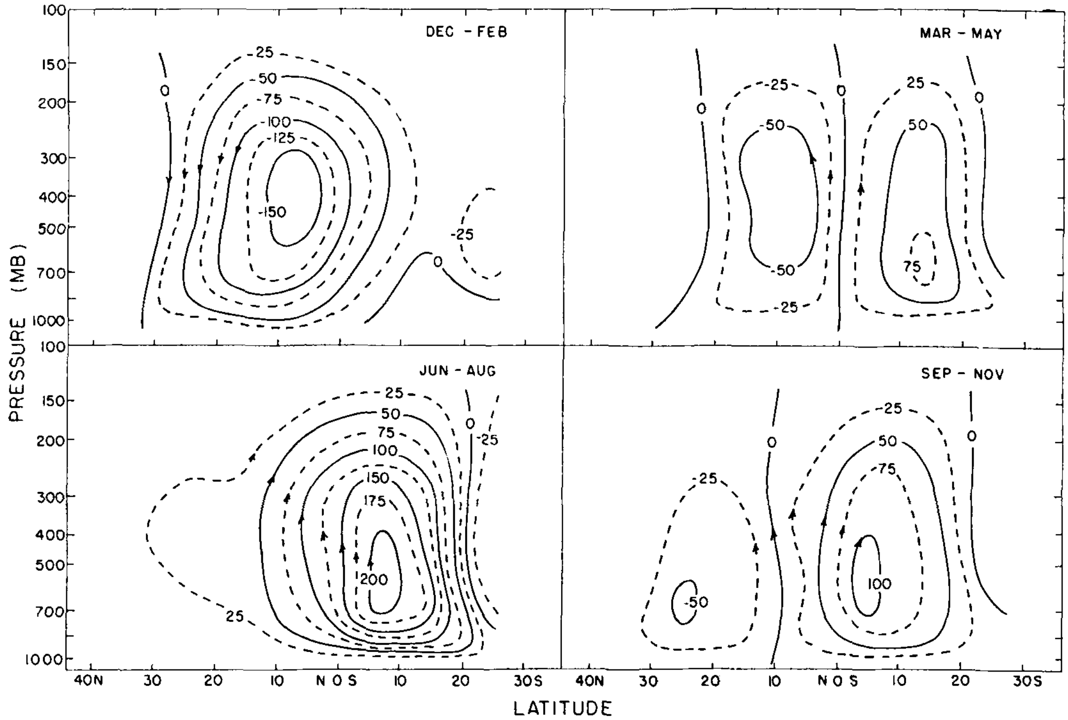

Fig. 1. Mass flux in meridional circulation. Units: 10^{14} g sec^{-1}.

With typical horizontal speeds involved in the upper and lower branches of the Hadley circulation of about 2 m sec^{-1} or 30° of latitude in two weeks, and vertical speeds of 5 mm sec^{-1} or 0.5 km day^{-1}, it can be seen that some of the air—but not a large fraction—could in principle pass through the cell and return to its hemisphere of origin (ignoring here any dynamic constraints). Nevertheless some mixing of a trace substance would occur while the air was absent from the hemisphere of origin and the above figure is probably only a small overestimate on this count. The tropospheric mass involved may be taken as the hemispheric area times 900 g cm^{-2} or 2.3×10^{21} g. The time necessary to exchange the whole mass at this rate is thus 320 days. This compares with estimates by Junge (1962*a*) of at least 400 days from CO_2 data and of about 4 years from CH_3T data, the former being given with the greatest confidence.

Transfer by eddy motions

If it is assumed that a measure of the transfer by large-scale eddy motions is given by the standard deviation of the meridional component of the wind then a cross-section representing the transfer can be simply constructed. Fig. 2 contains data for four three-month seasons; largest values of interhemispheric transfer occur near 150–200 mb with extremely low values in the lower troposphere. The latitude of the minimum transfer in the 150–200 mb layer follows the sun; smallest values accompany the region of strong rising motion in the Hadley cells. In the stratosphere the lateral transfer is a minimum in the 30–50 mb layer and smallest at latitudes of 10–20° rather than on the equator. The absolute minimum [as noted previously (Newell, Wallace & Mahoney 1966)] occurs near to the summer pole. The relationship between the data of Fig. 2 and the lateral mixing property of the atmosphere is neither clear cut nor simple. Considerable work is presently in progress by Kao and his coworkers (e.g. Kao, 1962, 1965; Kao & Bullock, 1964; Kao & Al-Gain, 1968). At the present stage of development the only approach available is to specify the spreading of marked particles in terms of a Fickian diffusion theory with

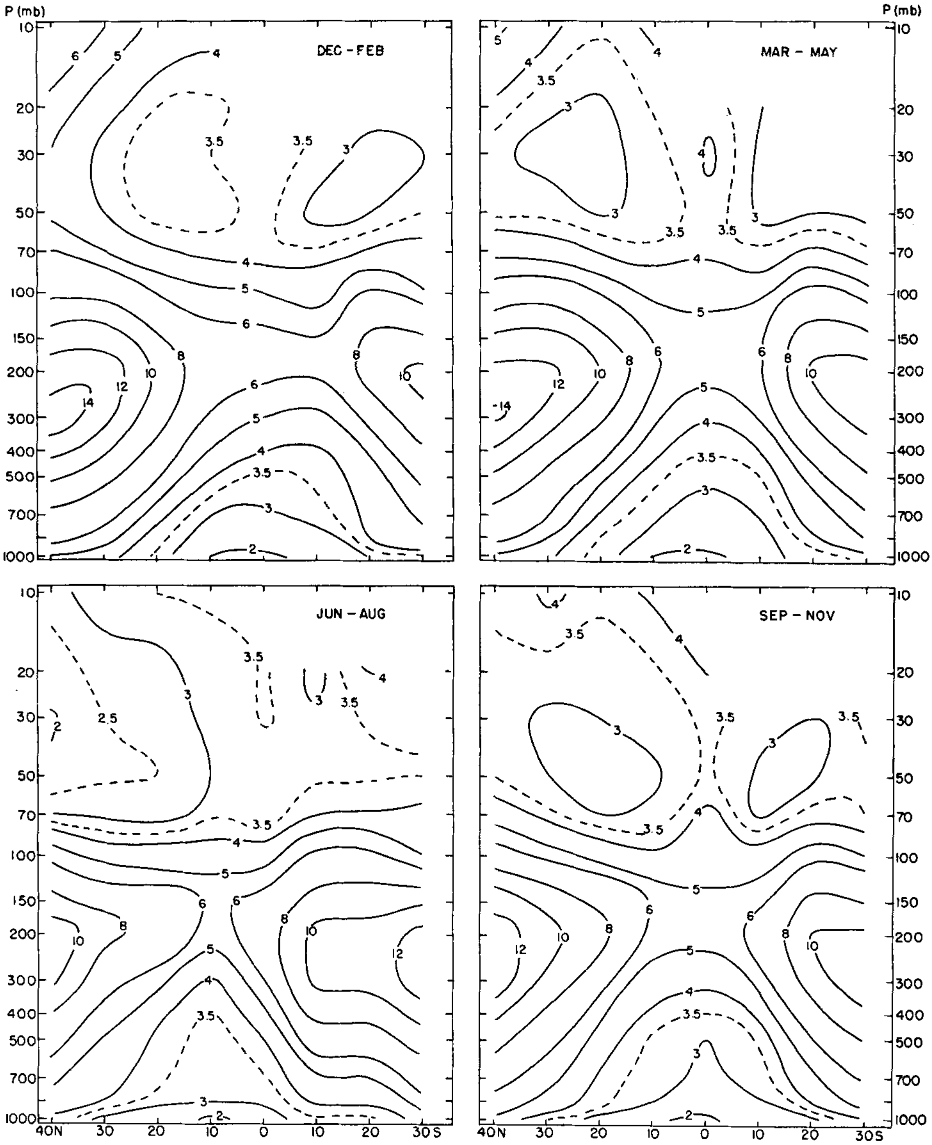

Fig. 2. Standard deviation of meridional component of wind. Units: m sec^{-1}.

K proportional to the variance of the meridional wind component (see references by Kao, loc. cit. and also Flohn, 1961). This approach was used to present K's from our meteorological results at the Utrecht CACR Symposium (Newell, 1963*a*). As an example suppose that the data of Fig. 2 can be converted to K values with the normalization factor obtained by setting $\sigma(v) = 15$ m sec^{-1} ($\overline{v'^2} = 225$ m^2 sec^{-2}) as corresponding to $K = 6 \times 10^{10}$ cm^2 sec^{-1}. (A value often used for middle latitudes at jet stream levels). So proceeding, values of K appropriate to the equatorial upper and lower troposphere and the tropical lower stratosphere are shown in Table 1. For comparison purposes a scale size of 1000 km is selected to obtain a relative measure of the speed of the diffusion process which is shown in the last column. When these speeds are compared with those of the mean meridional motions discussed earlier it may be

Table 1. *Typical eddy diffusion coefficients*

Region	$\sigma(v)$ m sec^{-1}	$\overline{v'^2}$ m^2 sec^{-2}	K cm^2 sec^{-1}	$\lvert v \rvert$ cm sec^{-1}
Upper troposphere Middle latitudes	15	225	6×10^{10}	600
Upper troposphere Equator	5.5	30	8×10^{9}	80
Lower troposphere Equator	2.0	4	1.1×10^{9}	11
Stratosphere 50 mb, 10–20° latitude	2.8	7.8	2.1×10^{9}	21

seen that the eddy activity plays a minor role in the exchange processes in the lower troposphere but is very important in the equatorial upper troposphere. In the northern hemisphere summer it will act to reinforce the transfer from the northern to the southern hemisphere for a source in the upper troposphere; in the northern hemisphere winter it will tend to provide mixing of material from the northern hemisphere into the southern hemisphere against the effects of the mean motion. With the values quoted a situation of almost zero net transfer could result for material in the northern hemisphere upper troposphere and of course an amplification of the transfer rate for material originating in the southern hemisphere in the same season. Table 1 also demonstrates that the transfer rate by eddies in the lower stratosphere at 50 mb, 10–20° latitude is comparable to that by mean meridional motions at the same place (cf. Newell & Miller, 1965; Vincent, 1968) underlining again the importance of efforts to deduce these mean motions. In middle and high latitudes the atmosphere may be considered to be well mixed in the vertical. Appropriate variance data for the vertical motions have been presented elsewhere (Newell, 1963*a*; Newell & Miller, 1968).

Comparison with trace substance distributions

In order to apply these meteorological results on both mean and eddy motions to make deductions about trace substance distributions, information on the mixing ratio gradients is required. It does not suffice to idealize the situation into a two box model with a transfer rate k given as

$$\frac{dC_2}{dt} = k(C_1 - C_2) \tag{1}$$

where C_1 and C_2 are the two hemispheric concentrations, although this form may be used as a first approximation to events. The strong dependence of cross-equatorial flows on season suggests that a more local space-time application of the equation is needed and this is presently precluded by the lack of the necessary observations.

Carbon dioxide has its source mainly in the northern hemisphere and global observations provide information about interhemispheric exchange rates. Bolin & Keeling (1963) presented global CO_2 data and recently a further attempt to model the observations has been made by Junge & Czeplak (1968). One interesting aspect of the observations is the phase change of the monthly maximum with the latitude of the maxima occurring in May at 50° N, July at the equator and August at 90° S. If the phase delay reflects interhemispheric exchange as suggested by the above investigators then the rapid transfer in the northern hemisphere summer is seen to be compatible with the reinforcement of mean and eddy motions at the same time as already noted.

Junge's original application of equation (1) to this situation gave an exchange time of at least 400 days with the k value assumed constant. If the equation is applied locally in time to the March–May period then the appropriate concentration difference between the hemispheres is 3.4 ppm (from Bolin and Keeling's table, 32° N–30° S) and the rate of change at 30° S is about 1 ppm/month. The k value obtained directly would correspond to a time of 100 days but of course source-sink terms are excluded. This time is shorter than the long term average we computed as 320 days for mean motions only.

There has long been a debate concerning the July maximum of surface ozone in the Antarctic which occurs in addition to the spring maximum which characterizes the northern hemisphere stations (Aldaz, 1965). It seems possible that ozone, with its much longer tropospheric residence time than particulates (4 months c.p. 1 month, Junge 1962*b*), can also cross the equator and reach the Antarctic. On this view the ozone exchanged between the northern hemisphere stratosphere and troposphere in spring is the source of the Antarctic July maximum. This ozone is introduced into a particularly favorable position (the sub-tropical upper troposphere) to participate in interhemispheric exchange according to the data presented above. It appears in the upper troposphere in late spring just at the time when the mean circulation from the northern hemisphere to the southern hemisphere is commencing in this region; this circulation continues through the summer and is reinforced by the eddy transfer. This potential mechanism seems to lend weight to the suggestion by Aldaz that there are two sources for ozone in the Antarctic: they are the northern hemisphere stratosphere, responsible for the July maximum, and the southern hemisphere stratosphere, responsible for the November maximum. Again the appropriate exchange rate is not the long term value deduced earlier but rather that appropriate to the particular season. A value of 1.4×10^{14} g sec^{-1} by the mean circulation and a like amount by the eddies seems reasonable for this region; the corresponding characteristic time is then about 100 days. This is close to the residence time for ozone in the troposphere (Junge, 1962*b*). There are a number of reasons for expecting the total ozone and total ozone transferred from the stratosphere to the troposphere to be smaller in the southern hemisphere than in the northern hemisphere (Newell, 1964). In addition the mean meridional motions oppose any flow from the southern hemisphere to the northern in the upper troposphere in the September–November period. Nevertheless such transfer would be expected to proceed in December and be visible at subtropical stations such as Mauna Loa (19° N). The one year's data available from this station does show an increase beginning in November. With the southern hemisphere spring maximum occurring later than that in the northern hemisphere and the possibility of transfer in November from the northern hemisphere stratosphere into the troposphere, it is clear that many more stations are needed over a long period before mixing ratio gradients between the hemispheres and within each hemisphere can be checked. Several surface ozone stations at low and middle attitudes in the southern hemisphere would be particularly valuable.

Table 1 also indicates that in the lower troposphere the eddies play a relatively minor role in cross-equatorial transfer. In summer the figures indicate little transfer is to be expected from the northern hemisphere into the southern hemisphere. The measurements of surface air radioactivity along the 80th meridian show evidence of such blocked transfer with very steep gradients at low latitudes (see e.g. Newell, 1963*b*; the gradient changes in the summers of 1959 and 1960 at the 10° N latitude, which roughly divides the southwards and northwards flows in the mean Hadley cells, are particularly evident). The actual interhemispheric transfer of such material in the upper troposphere would be expected to be small as the lifetime of particulates within the troposphere is about one month compared with 3–4 months for ozone.

If these views are correct trace substance distributions as functions of latitude should be significantly different in the lower layers from those at heights corresponding to 150 mb. In the former case mean motions will dominate within 20° of the equator whereas in the latter case both types of motion will be important and their resultant influence will be the sum of both effects.

Tropospheric-stratospheric transfer in the tropics

Rising motion in the direct Hadley cells in the tropics extends up to the tropopause and into the lower stratosphere. It is difficult to obtain its magnitude directly from the meteorological observations and in our previous work (Kidson *et al.*, *loc. cit.*) it was set equal to zero at 1000 and 100 mb in order to obtain the best possible estimates for the intervening layer. The heat budget at the tropical tropopause actually requires ascent and concomitant adiabatic cooling to offset the long wave radiative heating which constantly tends to erode the temperature minimum. Kennedy (1964) has demonstrated that the main contribution to this heating comes from carbon dioxide. From the heating rates computed by Rodgers (1967) (close to 0.25°K/day in the months of January, April, July and October) and the cooling by eddy flux divergence (Kidson *et al.*, *loc. cit.*) it is simple to estimate that the rising motion within 20° of the equator averages about 0.02 cm sec^{-1}. The eddy flux divergence cooling rates at the tropopause level are always smaller than the radiative heating rates. The ascent corresponds to about 20 m day^{-1} and as the horizontal speeds in the upper troposphere are typically 2 m sec^{-1} or about 2° latitude per day it can be seen that air may pass between hemispheres without being removed into the stratosphere. This rising motion thus has little effect on the interhemispheric exchange rates.

The rising motion is of interest in trace substance problems from another viewpoint, namely in that it serves as a source of tropospheric material for the stratosphere as originally propounded by Brewer (1949) for the case of water vapor. Mastenbrook (1968) has recently presented details of water vapor concentration measurement in the lower stratosphere. When the tropopause temperature is low the mixing ratios at higher latitudes are low as the mixing ratio is controlled by the limiting frost point at the tropopause. Mastenbrook interprets his data as indicating a low mixing ratio accompanying low tropopause temperatures (with the tropopause at a high altitude) in late winter. Our previous work indicates that the low temperatures actually span the equator and should be ascribed to the months December–May rather than to "winter" (Kidson *et al.*, *loc. cit.*). The northern hemispheric winter circulation seems to dominate conditions even at the equator. It is worth noting that the frost point measurements provide a limit to the latitudinal extent of the vertical motion penetrating into the stratosphere. A broad latitudinal region would encompass higher tropopause temperatures and concomitant higher frost points. The core of upward motion (from Fig. 2) penetrates further north in summer and it is then that the air passing into the stratosphere will be warmest.

If the assumption is made that the vertical motion occupies 30° of latitude centered on the equator with a value of 0.02 cm sec^{-1} then the appropriate mass flux is 2.6×10^{12} gm sec^{-1} (small compared with the interhemispheric mass flux). For a typical water vapor content of 2×10^{-6} g/g the water vapor source flux is 5.2×10^{6} gm sec^{-1}. Above 100 mb the stratosphere contains $100 \times 2 \times 10^{-6}$ g $cm^{-2} = 2 \times 10^{-4}$ g cm^{-2} of water vapor or a hemispheric total of about 5×10^{14} g. The appropriate replenishment time is then 10^{8} sec, or over 1000 days. This time is appropriate also for other gaseous trace substances entering the stratosphere by this mechanism. The return flow to the troposphere by transfer in middle latitudes has recently been discussed by Danielsen (1968).

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ОБМЕН МАССАМИ МЕЖДУ ПОЛУШАРИЯМИ ПО МЕТЕОРОЛОГИЧЕСКИМ И ТРАССЕРНЫМ НАБЛЮДЕНИЯМ

В обмене массами между полушариями играют роль два механизма: крупномасштабный вихревой перенос и средняя меридиональная циркуляция. На основе данных недавно законченных исследований общей циркуляции в тропиках даются оценки интенсивности обоих механизмов. Время долгопериодного обмена между полушариями найдено равным приблизительно 320 дням. Отсюда можно заключить, что когда вихревые процессы усиливают перенос путем средней меридиональной циркуляции, в верхней тропосфере в конце весны и летом может иметь место гораздо более быстрый обмен, который может быть ответственным за очевидную аномалию в сезонных вариациях концентрации озона на поверхности в южном полушарии. Наконец, поток массы между полушариями сравнивается с грубой оценкой потока массы от тропосферы к стратосфере, основанной на тех же самых тропических данных.