Transport of water vapour in a stratosphere–troposphere general circulation model. I: Fluxes

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SUMMARY

The fluxes of water vapour in the upper troposphere and lower stratosphere of a general circulation model have been calculated. Although the results are probably affected by some model characteristics, particularly the vertical resolution, they suggest that ingress of water vapour to the stratosphere may occur by large-scale eddy motions above the core of the subtropical jet stream, as well as by vertical motion through the tropical tropopause. The latter motion occurs predominantly in the equatorial western Pacific and Indian Ocean, and is connected with the monsoon circulation.

1. INTRODUCTION

The dryness of the stratosphere has been of interest since the first aircraft measurements in 1942 (Dobson et al. 1946); its importance for the physical interpretation of the meridional global-scale circulation was clearly stated by Brewer (1949). In recent years water vapour has been recognized as a key factor in the stratospheric photochemical balance, because it is the main source of the reactive OH free radical.

Despite many investigations with a variety of techniques there is still no certainty about the fundamental features of the vertical, latitudinal and longitudinal distribution of water vapour in the stratosphere: see for example Murgatroyd (1965), Harries (1976), Robinson (1980), WMO (1982). Current knowledge of the maintenance of this distribution has uncertainties even in a gross sense, and the detailed physical nature of the desiccation mechanism and cross-tropopause transports, particularly the influx to the stratosphere, is unknown.

It is therefore of interest to consider whether or not a general circulation model (GCM) can offer any insight on the global-scale transfer of water vapour across the tropopause. Such an approach has a number of inherent difficulties, among which are the need for long period integrations (the mass-weighted mean residence time of stratospheric air is 12–18 months) and the difficulties of representing well in a GCM such sub-grid-scale features as adiabatic drying by ascent in cumulus clouds and the structure of the tropopause. The latter points may be seen in the context of the model’s spatial resolution by reference to Fig. 1. Within these limitations, this paper seeks to examine the synthesis of the stratospheric water vapour distribution offered by a 514-day integration of the stratosphere–troposphere GCM described by O’Neill et al. (1982).

There were two major difficulties in comparing the model results with observations. The more important of these was the shortage of water vapour measurements in the upper troposphere and stratosphere; useful profiles measured over an annual cycle in fact exist at only four places on the earth’s surface. The second difficulty stemmed from the model’s behaviour; in both northern hemisphere winters simulated during the last 450 days of the run, there was a major stratospheric sudden warming. Accordingly, the model data should properly be compared with years in which the atmospheric data also have this characteristic. Unfortunately such a distinction is not available in the published water vapour records, although it might be possible to achieve such an analysis from the raw data. Dynamical analysis of the model sudden warmings and annual evolution may be found in O’Neill (1980) and O’Neill et al. (1982).
2. METHOD OF INVESTIGATION

The model used is derived from the 5-level model described by Corby and others (1972; 1977). For convenience, a brief description of the model is included here; for further details consult O’Neill et al. On each sigma surface ($\sigma =$ ratio of pressure to surface pressure) there are 4624 grid points distributed on sixty circles of latitude, which are three degrees apart. In the vertical the model uses ‘sigma coordinates’ with thirteen levels ranging approximately from 2 to 900 mb. The model has no explicit chemistry, but contains three tracers with empirically designed sources and sinks (water vapour, ozone and aircraft produced NO$_2$). Water vapour undergoes a full hydrological cycle including...
a sink due to condensation. There is, however, no source in the stratosphere to represent its production by the oxidation of methane.

The transport of water vapour occurs, on the larger scales resolved on the grid, by means of the advective terms on the l.h.s of Eq. (6) in Corby et al. (1972). In addition to this, there are two mechanisms which are present to preserve physical reasonableness in the solutions, and one to maintain numerical tractability. The dissipative behaviour of progressively smaller scales of motion for variable \( \chi \) must be represented for those scales not resolvable by the grid; it is achieved by use of a term \( K \nabla \cdot p_\ast \nabla \chi \) on the r.h.s. of Eq. (6) in Corby et al. For horizontal momentum and temperature, \( K \) is proportional to \( | \nabla \cdot p_\ast \nabla \chi | \); for water vapour it is fixed at \( 5 \times 10^{-4} \text{m}^2 \text{s}^{-1} \). This value is small, and was permitted by the use of a weak time-filter:

\[
\overline{\chi}^n = \chi^n + A(\chi^{n-1} - 2\chi^n + \chi^{n+1})
\]

where \( \overline{\chi}^n \) represents the time-filtered variable \( \chi \) at time \( n\Delta t \) and \( A = 5 \times 10^{-3} \). In addition to these two mechanisms, provision was made to correct negative mixing ratios, which may arise when the finite difference form of the continuity equation is applied to large gradients. If a negative value arose at a grid point, any advective transfer from that grid point was set to zero if it would have made the mixing ratio more negative. Such a term constitutes an implicit diffusion, as discussed by Mahlman and Moxim (1978); inspection of the frequency and magnitude of occurrences of negative mixing ratio suggests that it is smaller than the effects of the \( K \nabla \cdot p_\ast \nabla \chi \) term.

Heating of the atmosphere by shortwave radiation is simulated by using the climatological values for January and July; sinusoidal interpolation between these assumed extremes gives the variation throughout the year. The loss of heat by longwave radiation is parametrized by Newtonian cooling with coefficients derived from climatology.

The model was initialized by running it using conditions pertaining to January for 60 days; stratospheric water vapour had been set to zero at the start of this period. A simulation of the atmosphere was then made for a further 454 days using the heating and climatological sea surface temperatures varying in time through the annual cycle.

Data from this second period were analysed to investigate the manner in which water vapour was transported in the model. The fluxes of water vapour in both meridional and vertical directions were calculated as 5-day means at intervals of 45 days throughout the simulation. This period and interval were selected after examination of 90-day means and the monthly means for January and July, and the characteristics of 5-, 10- and 20-day averages, as suitable for studying the annual evolution. Following Lorenz (1967), these fluxes were decomposed into combinations of means and departures from these means of wind speed and concentration in both time and longitude. We may write this decomposition for the meridional flux as follows:

\[
[\rho \omega] = [\rho] [\bar{v}] + [\rho] [\bar{v}' v']^T + [\rho^* \bar{v}^*] + [\rho^* v^*].
\]

(i) (ii) (iii) (iv) (v)

Here \( \rho \) stands for the density of the tracer; \( v \) for meridional wind speed; \([a]\) denotes the zonal average of \( a; a^* = a - [a]; \bar{a} \) = the temporal, 5-day average of \( a; \) and \( a' = a - \bar{a} \). Thus, term (i) of (1) is the zonal mean flux averaged over time; (ii) is the meridional mean, stationary in time; (iii) the meridional mean, transient in time; (iv) the stationary eddy flux; and (v) the transient eddy flux. Analogous terms exist for the vertical transport.

The vertical velocity of the pressure surface as used in the model was converted to the Cartesian velocity, \( w \). Values of the air density, \( \rho \), were taken from the ICAO
standard atmosphere. This change in the vertical coordinate is consistent with the Cartesian vertical analogue of Eq. (1). Fluxes of mass may be analysed in an identical manner.

It should be noted that the data in the figures have been converted to use pressure as the vertical coordinate.

The time history of the zonally averaged distribution of water vapour has been investigated, as have the dynamic variables that influence it. The effects of temperature changes on the distribution of water vapour throughout the integration at specific latitudes have been investigated. Lastly, the global fields of the temporal average of the covariance between water mixing ratio and the components of velocity have been analysed.

3. Determination of Fluxes of Mass and Water Vapour

During the computation of the quantities shown in Eq. (1) twenty individual sets of data corresponding to 6-hour intervals were used in one case, but it was found that no important loss of information occurred by using only five data-sets at 24-hour intervals.

Both vertical and meridional fluxes, $F_v$ and $F_m$ respectively, were calculated, but for convenience they were combined into a single vector with length, $l$, and direction,

![Figure 2. Zonal mean fluxes of mass in the model latitude-height plane defined by equations (1), (2) and (3) in the text. Note that the plotted vertical coordinate is pressure; height was used in the calculations. (a) Term (i) of Eq. (1), $[\rho u]$, $[\rho w]$.](image-url)
\[ \theta = \text{angle made with horizontal, as shown below:} \]
\[ l = k_1 \log_{10}(F_m^2 + F_m^2)^{1/2} \]
\[ \theta = \tan^{-1}(k_2(F_m/F_v)). \]  

Here, \( k_1 \) is a constant of proportionality and \( k_2 \) a scaling factor to allow for the different vertical and meridional distances on the earth represented by unit lengths on the diagram. The logarithmic weighting when calculating \( l \) compensates for the rapid fall-off in the fluxes with height through their dependence on the density.

(a) Fluxes of mass

The total flux of mass as represented by term (i) of (1) is equivalent to the determination of the meridional streamfunction. Since the individual contributions to the total flux are not necessarily non-divergent, we are unable to calculate analogues to the streamfunctions for each component flux. Since the objective in this section is to determine the relative roles of the different modes of transport defined by Eq. (1), the streamfunctions are not presented in this work. These do appear, however, in O'Neill et al. (1982). They noted tropospheric cells corresponding to the Hadley, Ferrel and polar mean circulations which extend into the stratosphere. This does not necessarily indicate that air is passing from the troposphere through the tropopause, as these circulations do not constitute trajectories of parcels of air.

Figure 2(b). Term (ii), \([\rho] [v], [\rho] [w]\)
In this work, the decomposition of fluxes for the second January and for July will be considered. In the 5-day average at the beginning of January, the total flux of mass (Fig. 2(a)) is dominated by the stationary, zonal mean term as shown in Figs. 2(b) and 3. Only in the winter stratosphere does the contribution from the stationary eddies become appreciable (Fig. 3). The transient mean term has no discernible structure and is negligible in magnitude. The two eddy terms have similar structure, but the stationary term is larger than the transient (Fig. 3).

![Figure 3](image)

Figure 3. Model meridional mass transport at 70 mb, average for days 360–364 (January). Solid line, mean cell (term (ii) of Eq. (1)). Dashed line, stationary eddy (term (iv)). Dotted line, transient eddy (term (v)). Dash-dot line, net flow (term (i)). Term (iii) is too small to appear on the scale given. See Fig. 2 for note on vertical coordinate.

In July, numerical values of the fluxes are all smaller than their counterparts in January. Throughout the 100 mb surface the total flux is dominated by the stationary zonal mean term (Fig. 4). A major difference between the hemispheres is evident. In the stratosphere, the transport of mass by eddies is appreciable in the northern winter but not in the southern winter.

![Figure 4](image)

Figure 4. Model meridional mass transport at 100 mb, average for days 180–184 (July). Legend as in Fig. 3.
Figure 5. Zonal mean fluxes of water vapour in the model, defined by Eqs. (1), (2) and (3) in the text.

(a) Term (i) of Eq. (1), [q_v], [q_w].

(b) Fluxes of water vapour

The division of the water vapour flux into components is more complicated. Throughout the troposphere, the total flux is directed upwards and polewards (Fig. 5(a)). In the stratosphere, the direction of the flux follows that of mass. The stationary mean term (Fig. 5(b)) has a similar structure to that for mass, but this is to be expected considering that both terms consist of the product of the average component of velocity and the appropriate averaged density. The transient mean flux is distributed randomly in direction. The two eddy terms have regular structures and are similar to one another.

The transport of water vapour during January in the model’s lower troposphere is dominated in the tropics by the mean meridional circulation (Fig. 6(a)) although the eddies do play some part especially in the upper troposphere (Fig. 6(b)). In the extratropical troposphere, however, eddy motions are the most important mode of transport (Fig. 6(a)). The northward transport in the tropical stratosphere is determined by the mean meridional circulation (Fig. 6(d)) except in the lower stratosphere where both eddies and mean cell contribute to the northward component (Fig. 6(c)). This is also true in the extra-tropical latitudes with the eddy motions, and particularly standing eddies, dominating in both the northwards (Fig. 6(d)), and vertical (Fig. 6(e)) directions.
Figure 5(b). Term (ii), $\overline{\langle q \rangle [v]}$, $\overline{\langle q \rangle [w]}$. 

Figure 6. Model fluxes of water vapour, average for days 360-364 (January). Legend as in Fig. 3. (a) Meridional, 700 mb.
(b). Meridional, 200 mb.

(c). Meridional, 70 mb.

(d). Meridional, 30 mb.

(e). Vertical, 200 mb.

Figure 6, (b) to (e).
Figure 7. Some observational data on the fluxes of water vapour. (a) to (d): after Oort and Rasmussen (1971). (a) January, $\bar{[q]} \bar{[v]}$, g kg$^{-1}$m s$^{-1}$; (b) January, $[q'v']$, g kg$^{-1}$m s$^{-1}$; (c) January, $[q^*v^*]$, g kg$^{-1}$m s$^{-1}$; (d) January, $[q][v] + [q']v' + [q^*v^*]$, g kg$^{-1}$m s$^{-1}$ ($= [qv]$).
Figures 7(a) to (d) show observed meridional fluxes of water vapour due to the zonally averaged meridional circulation, the transient eddies, the standing eddies and the sum of the three for January, redrawn from Oort and Rasmusson (1971). This sum may be interpreted as the total flux. In the tropics, the mean meridional circulation provides the main contribution to this total flux, while in extra-tropical latitudes the eddies are the dominating mode of transport. These observations and the general structure of the flux components are simulated by the model, as may be seen by comparing Figs. 7(a) to (d) with 5(a) and with 6(a). Peixoto et al. (1978) computed the three main components of the meridional transport of water vapour for the whole troposphere during IGY. These components averaged for the period October to March are reproduced in Figs. 7(e) to (g). They are reasonably consistent with the data of Oort and Rasmusson in the northern hemisphere.

The domination of transport in the tropics by the mean meridional component is evident in both model results and observations. In the model the most important mode of transport in the extra-tropical latitudes of the southern hemisphere is the transient eddies, but the data of Peixoto et al. suggest that both the mean meridional circulation and the transient eddies will be important. This disagreement may be due to the difference between the average of Peixoto et al. over a 6-month period containing January and the average over a 5-day period in January as calculated by the model, although the latter was similar to the appropriate monthly and winter means. The sparse data coverage over the southern oceans could also contribute to the discrepancy.

For the stratosphere, comparisons of atmospheric data with model results are not possible due to the scarcity and low accuracy of water vapour observations at high levels.
Similarly, estimates of vertical fluxes are difficult even in the troposphere, since the vertical wind speed may only be inferred indirectly from observational data.

From the results of Oort and Rasmussen, it appears that the vertical transport by the mean meridional circulation is much greater throughout the troposphere than that by the stationary eddies. In the model this is true in the tropics, but in extra-tropical latitudes the eddy transports are more important. The structure of the total transport, however, does match the observations with upward transport in the tropics and mid-latitudes and downward transport in the sub-tropics.

In July a similar dependence of transport upon the horizontal and vertical eddy terms exists to that discussed above for January. However, the net transport of water vapour in the northern hemisphere is greater than that in the southern hemisphere in both months, suggesting that the eddy motions are stronger in the northern hemisphere. In the tropics, the model suggests that the net flux is directed northwards (Fig. 8), as opposed to southwards in January. From the data of Oort and Rasmussen it is clear that in July the northward transport by transient eddies is weaker than in January and that the maximum is shifted polewards. However, transports by stationary eddies are relatively more important in July than in January. In the tropics, the transport of water vapour by the mean meridional circulation is northwards in July as opposed to southwards in January, reflecting the seasonal change in direction of the returning branch of the Hadley cell. This observation is clearly evident in the model data, see Figs. 6(a) and 8. Throughout the integration, the tropospheric flux of water vapour is directed polewards and upwards while that in the stratosphere is more closely aligned with the flow of mass. The time history of the fluxes of mass is given in the streamfunctions of O’Neill et al. (1982) and will not be reproduced here. For certain latitudes the zonally averaged transport of water vapour in the troposphere moves in the opposite direction to that of mass. This is because ascending air tends to be more moist than descending air. It is interesting to note that throughout the model troposphere, the direction of the water flux is approximately parallel to the spatial gradient of water mixing ratio. In the stratosphere these gradients are smaller and mass and water fluxes are more aligned.

![Figure 8. Model meridional flux of water vapour at 700 mb, average for days 180-184. Legend as in Fig. 3.](image-url)
4. Analysis of zonally averaged distribution of water vapour

In the model water vapour at any level in excess of its value at saturation is removed as precipitation to the surface, providing that the relative humidity of the level below exceeds 80%. This modification of the original model, which had no such restriction, was found to give more realistic temperature profiles in the tropics (O'Neil et al. 1982).

Saturation is with respect to ice at temperatures <233 K, and w.r.t. water at temperatures >233 K. In the real atmosphere, the transition to glaciation may occur in the region 240–270 K, depending on cloud microphysics. In the absence of a reliable climatological distribution this simple formulation was adopted. It should not seriously affect our conclusions. Since fluctuations in temperature arise through the seasonal variation in the solar heating field, desiccation of the atmosphere through removal of excess water vapour can occur diabatically rather than relying upon adiabatic vertical motions to provide the low temperatures required. The field of temperature produced by the model includes a very cold winter pole and also minima in temperature at about 100 mb in the tropics and 200 mb in extra-tropical latitudes. The corresponding field of mixing ratio is clearly influenced by these temperatures, with the lowest mixing ratios coinciding with the minima in temperature.

Figure 9(a) shows the zonally averaged temperature field during the northern summer while Fig. 9(b) shows that for the second northern winter. The corresponding fields of water vapour mixing ratio are shown in Figs. 9(c) and (d).

Oort and Rasmusson (Figs. 10(a), (b)) show their data for January and July. The mixing ratio of water is a maximum at the surface in the tropics and decreases with both
Figure 9(b). Temperature, day 405 (February), K.

Figure 9(c). Water vapour, day 190, mass mixing ratio $\times 10^6$. 
Figure 9(d). Water vapour, day 405, mass mixing ratio ×10^6.

Figure 10. Observed zonal mean fields of water vapour, after Oort and Rasmusson (1971). (a) January, mass mixing ratio ×10^6, (b) July, mass mixing ratio ×10^7.
height and latitude. The rate of decrease is larger in January than in July. The same pattern is found in the model (Fig. 9(c)) but its absolute values of the mixing ratio are somewhat less below 700 mb.

Figure 11(a) shows the change of temperature with height and time at high latitudes in the northern hemisphere of the model. The stratosphere is cold in winter and warm in summer as previously observed in Figs. 9(a), (b). Figure 11(b) shows the corresponding distribution for water vapour mixing ratio. Comparison of these figures shows a clear correlation between temperature and mixing ratio with the qualification that variations in water vapour seem to lag those in temperature, especially at high altitudes.

At high latitudes, the stratosphere in the second northern winter is appreciably warmer than the first, perhaps because of the long period of cooling before the first winter during the initialization. The distribution of water vapour also shows higher values
in the stratosphere of the second winter than in the first, but has a drier region at 200 mb.

During the 450-day integration, there is an upward trend in the stratospheric water vapour content above about 70 mb which is visible in Figs. 11(b) and 12(b). However, its amplitude is small compared to the annual variation below 50 mb, and the lack of equilibration should not have a large effect on our analysis, which concentrates on levels near the tropopause.

![Figure 12](image)

Figure 12. Model altitude–time sections, zonal mean at 10.5°N. (a) Temperature, K. (b) Water vapour, mass mixing ratio \times 10^9.

Figure 12(a) shows that in subtropical latitudes of the northern hemisphere the minimum temperature occurs at 100 mb in early spring while a maximum is observed in early autumn. The amplitude of this annual variation is less than that seen in high latitudes. A similar variation may be noted in the water vapour field (Fig. 12(b)) although there is evidence to suggest that the phase of the water distribution lags that of temperature by about three months at 100 mb. The phase of the water distribution also changes with
height. The maximum at 50 mb occurs about one month after it appeared at 100 mb. The phase of the temperature variation remains constant with height. These findings suggest that the lag in water variations at high levels is due to the time taken for the parcels of air with water vapour mixing ratios reduced by the lower temperatures at 100 mb to arrive at higher levels. It may also be noted that although there is little inter-annual variation in temperature at subtropical latitudes, the water vapour mixing ratio at higher levels tends to be higher in the second winter than in the first, indicating the long times needed to equilibrate there. We may also compare the annual variation of stratospheric water vapour at Washington D.C., U.S.A., as measured by Mastenbrook and Daniels (1980) (Fig. 13(a)) with the variation at the grid point corresponding to the position of Washington in the model for the last year of the integration (Fig. 13(b)).

The data of Mastenbrook and Daniels show a maximum in the lower stratosphere

![Diagram](image-url)

Figure 13. Altitude–time sections of water vapour at or near Washington, D.C.
(a) Observations, mass mixing ratio \( \times 10^8 \) (Mastenbrook and Daniels 1980).
(b) Model, mass mixing ratio \( \times 10^9 \), nearest grid point.
in late summer, while in the model, the maximum occurs earlier. At 100 mb, however, the maximum occurs a little later in the model. The observed minimum in mixing ratio at about 170 mb in the winter and spring is not reproduced. The observed increase above this altitude may be due to the oxidation of methane, which was not represented in the model. Generally, at these lower stratospheric altitudes the simulated mixing ratios are higher than in the real atmosphere. However, the steeper gradient between 200 and 100 mb compared with that between 100 and 50 mb is represented qualitatively.

5. COMPARISON OF FIELDS OF WATER VAPOUR WITH OTHER MEASUREMENTS

Murumatsu (1981) describes measurements of water vapour mass mixing ratios measured over Japan. Figure 14(a) shows his data for the winter period. Figure 14(b) shows the mixing ratio of water averaged over 5 days in the second January simulated by the model along 140°E. The model data show a minimum in mass mixing ratio at about 100 mb and south of 30°N corresponding to the low temperatures at the tropical tropopause, and a second minimum between 60 and 50°N, apparently due to the low temperatures in the northern polar regions during the polar night. Murumatsu’s data show a similar distribution but with mass mixing ratios much lower than those simulated by the model.

For example, at about 45°N over Japan, Murumatsu observed mass mixing ratios of less than $1.5 \times 10^{-6}$ between 174 and 100 mb. At 100 mb the temperatures required to obtain this mixing ratio by desiccation would be about $-85$ °C. The monthly mean temperatures for January above Japan at 100 mb (Goldie et al. 1958) range from $-48$ °C in the north to $-64$ °C in the south. These temperatures are too high to dry the atmosphere in situ to the extent noted by Murumatsu. Goldie et al. noted a maximum in temperature of about $-46$ °C to the north-west of Japan in their average for January. This feature may correspond to an area of systematic downward motion, hinted at in Fig. 16(a), perhaps related to the strong jet stream over Japan. This is consistent with
the discussion by Manabe and Mahlman (1976) of their Fig. 10.4. Assuming that the air becomes drier with altitude, the region of descending air would be drier than its surroundings. Model trajectories discussed in part II of this paper also suggest possible explanations of Murumatsu’s data.

Another possible cause of the low values of mixing ratio observed is through horizontal advection from the northern polar regions. If the temperatures were low enough to reduce the mixing ratios near the pole, such motions would produce a field of mixing ratio similar to that observed. According to Goldie et al., the lowest mean temperature in these regions at 100 mb during January is about \(-64^\circ\text{C}\). This is too warm to produce the desiccation necessary to reduce the mass mixing ratio to \(1.5 \times 10^{-6}\). The low values of mixing ratio noted by Murumatsu may not be explained by considerations of the 100 mb temperature field outside the tropics. However, recent satellite observations using the SAM II instrument (McCormick 1981) have suggested the occurrence during the arctic winter of stratospheric clouds, well correlated in space and time with sufficiently low temperatures for ice crystal formation. Such data give a wider context to earlier observations of nacreous clouds. Provided that the ice crystals survive long enough to fall or be transported to significantly lower altitudes, such data may offer an explanation of the decrease in water vapour mixing ratio observed in the lower stratosphere between Tateno (36°N) and Sapporo (43°N). The air dried by loss of ice crystals would have to be moved equatorwards and downwards.

Because of the absence of any representation of methane oxidation, and the comparatively short integration time from the initial conditions of zero stratospheric water vapour, comparison with the scattered, mainly mid-latitude, balloon profiles of water vapour reaching to the stratosphere above 30 mb would not be useful.

6. ZONAL AVERAGE OF SINKS OF WATER VAPOUR

The zonally averaged sources and sinks (not shown) indicate that water was removed from most latitude bands throughout the troposphere. The main sources of vapour were in the lower subtropical troposphere. The latitudinal distribution of net precipitation in the model is similar to that computed by Peixoto (1970) for the real atmosphere.

7. THE FIELDS OF COVARIANCE BETWEEN WATER VAPOUR AND VELOCITY COMPONENTS

The covariances of water vapour, averaged in time, and the components of velocity were computed from data from days 360 to 370 of the integration. These were compared with the average mixing ratio distribution for the same period to examine the relationships between the tracer and velocity fields.

In the tropics there is a positive correlation between the vertical flux of water and its mixing ratio (Figs. 15(a), (b)). This reflects the fact that upward moving air tends to have a higher mixing ratio than downward moving air. Also, the magnitude of the vertical flux in the tropics is larger than elsewhere. This is due to the strong tropical convection carrying air of high mixing ratio, and is in agreement with the observed structure shown by Bannon and Steele (1960). Observations of the real atmosphere by Bjerknes (1969), Johnson and Townsend (1981) and Lorenc (1982), among others, have indicated that these vertical motions have a consistent longitudinal structure. From Fig. 15(a) it is noticeable that for the model in January, the largest vertical fluxes seem to occur close to the sub-solar latitude and at the eastern margins of the continents of the southern hemisphere. These positions correspond to the low-level maxima in isentropic
Figure 15. Model data at 500 mb, average of days 360–370 (January). (a) Product of water vapour concentration and vertical velocity.

Figure 15(b). Water vapour mass mixing ratio ×10⁶.
mass transport for the same time presented by Johnson and Townsend and which indicate ascending air. Figures 16(a), (b) show the divergent winds at 100 mb and 850 mb respectively for January deduced from the FGGE data by Lorenc (1982). They show *inter alia* that the air to the north-east of Australia was ascending. This motion pattern seen in both atmosphere and model indicates that the monsoon circulation, which is the horizontal
Figure 17. Model meridional fluxes of water vapour, average of days 360–370. Closer shading, northward fluxes; wider shading, southward fluxes. (a) 900 mb.

Figure 17(b). 100 mb.
manifestation of the vertical motions implied by Johnson and Townsend and by Lorenc, is also simulated in the GCM; it has been discussed by Corby et al. (1977). In January the low-level flow over the Indian Ocean and Indo-China is equatorward, thereby producing the negative meridional flux of water vapour seen in Fig. 17(a). Conversely, at higher levels the flow and hence the flux are in the opposite direction, as shown in Fig. 17(b).

In extra-tropical regions, the longitudinal distribution of mixing ratio seems to be determined by the meridional motions, shown in Fig. 17(a). Poleward motions tend to transport more water than do equatorward motions due to the latitudinal gradient in water vapour caused by convection in the tropics and desiccation by the low temperatures in the polar night. In most areas there is a correlation between poleward and upward fluxes. Since the surface of potential temperature in the extra-tropical stratosphere tends to slope in the opposite sense, it would seem that the motions of water vapour at these levels are across the isentropic surfaces. This non-adiabatic behaviour is to be expected when it is noted that removal of water vapour by condensation occurs throughout the model troposphere.

At levels above 100 mb the characteristic wavelength of longitudinal variations is longer than for tropospheric levels, as shown in Figs. 18(a), (c). Maximum mixing ratios are found in mid-latitudes with minima in equatorial and polar regions (Fig. 18(b)). There are slight indications that the water vapour mixing ratio throughout the atmosphere at these levels is influenced by vertical motions, but the correlation seen in the troposphere is not found.

![Figure 18](image_url)

**Figure 18.** Model data at 70 mb, average of days 360-370 (January). (a) Vertical flux of water vapour; closer shading, upward; wider shading, downward.
FLUXES OF WATER VAPOUR

Figure 18(b). Water vapour mass mixing ratio.

Figure 18(c). Meridional flux of water vapour; closer shading, northward; wider shading, southward.
8. AN ANALYSIS OF FLOWS BETWEEN TROPOSPHERE AND STRATOSPHERE

The position of the tropopause in the model is difficult to determine because of the comparatively low vertical resolution. From meridional sections of the temperature field, however, the tropopause appears to be situated at about 200 mb in extra-tropical latitudes and 100 mb in the tropics. From the values of the vertical and meridional fluxes of mass and water, the amount of each passing through this so-defined ‘tropopause’ in a certain time may be computed. The mass of air passing through the ‘tropopause’ from troposphere to stratosphere was computed at various times throughout the experiment. The results

Figure 19. Vertical mass flow through model ‘tropopause’ (see text). Positive ordinate indicates flow from troposphere to stratosphere.
(b) Global rate of change of mass.
(c) Model flux of water vapour across the tropopause. Solid line, vertical flow through horizontal sections of tropopause (dot–dash line). The dashed line indicates total flow, including horizontal flow through the vertical sections of the tropopause. Units, $10^9\text{kg s}^{-1}(3^\circ\text{latitude})^{-1}$. Data are the mean of days 360–364 (January).
are shown in Figs. 19(a), (b). The flow of mass through the 'tropopause' in each hemisphere exhibits a clear annual variation, with a maximum influx to the stratosphere in late spring–early summer and a maximum efflux during winter. The changes in the flow of mass through the northern hemisphere 'tropopause' are largely balanced by changes in the opposite sense in that of the southern hemisphere, reflecting the fact that the total mass above a given pressure surface must remain constant. This implies that in the model stratosphere there is a transfer of mass from the summer hemisphere to the winter hemisphere. Danielsen and Mohnen (1977) detected an annual cycle in the mass of air moving from the stratosphere to troposphere in the northern hemisphere by means of folds in the tropopause. From the data of Roe (1981) for the monthly variation of mass in the northern stratosphere, the rate of change of stratospheric mass in the northern hemisphere may be computed. Comparing these data with those from the northern hemisphere in the model, it is found that the phase of the cycle in the model precedes that observed in the atmosphere by about four months. However, since in the atmosphere there is an annual variation in the position of the tropopause, the two sets of data are not directly comparable. In addition, it is apparent that the observed rate of change is in phase with the contribution due to tropopause folding. The folding of the tropopause can only describe the exit of mass from the stratosphere, and in the model this process is not readily quantifiable, due to the lack of resolution in the vertical and the difficulty of producing an objective scheme to analyse folding events over the entire space-time domain.

Figure 19(c) shows the average rate of flow of water vapour across the model tropopause during the second January. Most of the water vapour crosses in mid-latitudes, perhaps suggesting that vertical transport is more efficient near the baroclinic zone. The large 'spikes' situated at the 'breaks' in the tropopause are due to the meridional flux through the vertical sections of the model tropopause. However the tropopause is defined, it is clear that these meridional motions transport an appreciable quantity of water vapour from the upper tropical troposphere to the lower mid-latitude stratosphere. The way in which these flows are manifested in the horizontal plane may be seen by analysing the covariance of velocity components and mixing ratio, averaged over a certain time.

Figure 17(b) shows the covariance between the meridional component of velocity and water mixing ratio at 100 mb, for the first part of the second simulated January. The pattern at 200 mb is similar and the structure of 100 mb is, therefore, consistent with that for the whole vertical section between troposphere and stratosphere. In this region, the tropical latitudes are dominated by regions of poleward flux, indicating that water vapour is moved from troposphere to stratosphere approximately in the horizontal plane. Figures 6(b) and (e) indicate that the motions at subtropical latitudes are dominated by eddies.

Figure 18(c) shows that for tropical northern latitudes at least, the flow at 70 mb is dominated by equatorward flow into the dry region above the tropical tropopause (Fig. 18(b)). This probably accounts for the difference in the zonally averaged fields of mixing ratio at these altitudes between July (Fig. 9(c)) and February (Fig. 9(d)). Figure 6(c) shows that at 70 mb the meridional flux of water in the northern tropics is dominated by eddies. At higher altitudes (Figure 6(d)) this is also true for the equatorward flow in northern mid-latitudes, where the standing eddies are particularly important.

It seems, therefore, that the schematic circulation deduced by Manabe et al. (1965) from their GCM is reproduced in this model (Figs. 5 and 6), although we also find a meridional eddy flux from subtropical upper troposphere to lower mid-latitude strato-
sphere. They also noted high mixing ratios in subtropical latitudes which they attributed to large-scale eddies in mid-latitudes and the meridional circulation in the tropics. In the model considered in this work, the air immediately above the tropical tropopause is very dry, and hence any enhancement of mixing ratios in the subtropics must come from below. To summarize, these high mixing ratios seem to be transported from the troposphere by eddies in both horizontal and vertical planes.

9. **Vertical fluxes in the tropical troposphere**

In a previous section it has been shown that the monsoon circulation played a large part in determining the distribution of water vapour in the tropical troposphere of the Indian Ocean. According to Brewer's (1949) hypothesis, the mixing ratio of water vapour in the stratosphere is determined by the temperature of the rising air at the tropical tropopause. He envisaged that transport by a meridional cell would carry air upwards through the tropical tropopause, polewards and finally downwards over the pole. This circulation is qualitatively the same as that required to balance the net heating due to the latitudinal distribution of radiative sources and sinks calculated by Murgatroyd and Singleton (1961), and is also similar to the mean meridional circulation in isentropic coordinates computed by Townsend and Johnson (1981). Dunkerton (1978) identified it as the Lagrangian circulation as determined approximately by the Eulerian mean field of diabatic heating, while Pyle and Rogers (1980) produced a similar flow from the differences between the Eulerian circulation and that induced by steady, non-dissipating waves.

This circulation is not zonally symmetric in the atmosphere. In January 1979 ascent was favoured near the southern continents, as shown by Johnson and Townsend (1981). Newell (1979) also shows schematically a similar zonal circulation. In July 1979 the centres of ascent were located in Central America and South-East Asia. This longitudinal manifestation of the Hadley cell in the Pacific sector has been termed the Walker circulation by Bjerknes (1969). Analogous cells exist for other longitudes, related in phase to the Walker circulation. Stoeckenius (1981) shows that the inter-annual variations in tropical precipitation may be explained by similar variations in the intensity of the zonal cells, but the intensity and position of the Hadley cell, which is effectively the zonal average of these zonal circulations, are relatively unimportant. The variations in intensity of the zonal cells will presumably affect the temperature and altitude of the

![Figure 20. Observed 100 mb temperatures and winds, 1979, plotted from tabulated monthly mean radiosonde data (NOAA 1980). (a) January. (b) July.](image-url)
tropical tropopause and hence cause similar variations in the mixing ratio of water vapour in the lower stratosphere.

Figures 20(a) and (b) show in diagrammatic form the distribution of temperature in the tropics at 100 mb from measurements by the radiosonde network during 1979 (NOAA 1980). Despite the variety of sondes employed and the variable number of ascents per month, the data can be drawn up into coherent patterns. The lowest temperatures were found in the same locations as the main areas of ascent determined by Johnson and Townsend. For saturated air, the lowest temperatures observed correspond at this level to water vapour mass mixing ratios of about $2 \times 10^{-6}$. Large regions of the tropics would have corresponding values of less than $4 \times 10^{-6}$. If the temperatures of ascending air in the tropics determine the extent of saturation in air parcels entering the stratosphere, and if the desiccation process removes water vapour in excess of the amount at saturation, then, neglecting the source of water due to the oxidation of methane, the mixing ratio of water measured in the stratosphere at mid-latitudes cannot exceed that above the tropical tropopause. The minimum temperatures at 100 mb in July are higher than those in January despite the larger mass convergences at low levels observed in July by Johnson and Townsend. The location of the minimum in temperature is the same as the centre of ascending air deduced by Johnson and Townsend. Both the

![Diagram](image)
patterns in temperature and their change with season are consistent with the data of Goldie et al. (1958). These observations and conclusions are similar to those made recently by Newell and Gould-Stewart (1981) who used data from IGY.

The corresponding temperatures simulated by the model (Figs. 21(a) and (b)) are higher than those observed, which is consistent with the higher mixing ratios found in the lower stratosphere of the GCM. However, the spatial distribution of the extremes of temperature tends to follow that of the atmosphere.

In January, the simulated mass mixing ratio at 100 mb in the tropics (Fig. 21(c)) is correlated with the temperature at the same level (Fig. 21(a)), although the spatial structures of the two variables do not match exactly. In particular the low values of mixing ratio found in the Indonesian sector tend to correspond to the low temperatures in the same region; the minimum in q found at 70 mb (Fig. 18(b)) shows this more clearly. This feature is also compatible with the region of low ozone column density observed in both the real atmosphere and in the model, as shown in Figs. 1(a) and (b) of Allam et al. (1981). In July the mixing ratios show little correlation with temperature, as shown in Figs. 21(b) and (d). At this time the spatial structures of the two variables are similar, but shifted in longitude relative to each other. The minimum in water vapour is situated about ninety degrees to the west of that in temperature.

In the model tropics the magnitudes of the mixing ratios are not at first sight compatible with the temperatures. The lowest spot temperatures found at the model 100 mb level in the tropics are about 192 K, which corresponds to a saturation mass mixing ratio of about $3 \times 10^{-6}$. The lowest mixing ratios found in the model are about one third this value. This discrepancy may be partly explained by an examination of the finite difference equations for condensation during the convective process (Corby et al. 1972, Eq. (41)). The temperature used to calculate the saturated mixing ratio at a certain level is derived from the temperature at the 'half level' below, which is calculated using a lapse rate interpolation. Precipitation was not allowed unless the grid point below had a relative humidity of at least 80%, as a crude way of allowing for evaporation from falling hydrometeors. Figure 22 shows the vertical distribution of temperature for a particular column for which the mass mixing ratio of water at 100 mb was $1 \times 10^{-6}$. At each level the convective scheme's effective condensation temperature at which the saturated mixing ratio was calculated was lower than the ambient temperature at the same level, by about 3 K. If applied at 100 mb this yields a mixing ratio of $3 \times 10^{-6}$, if

![Figure 22. Tests of the saturation mixing ratio and temperature profiles in the model. Data are for day 360 (4.5°N, 90°E). See text for discussion](image-url)
applied at the 'half level' between 200 and 100 mb the value is $2 \times 10^{-6}$. However, even these reduced temperatures remain too high to account for the very lowest mixing ratios found at the 100 and 200 mb model levels. The above analysis is complicated, however, by the effects of horizontal advection and mixing, which change with every 10-minute time step, while the model fields are available at 6-hour intervals. The lowest mixing ratios, which are found only at a low percentage of the total number of grid points, could also be due to truncation error in the vertical advection scheme arising from the steep gradient of mixing ratio in the vertical.

The relative humidity at different pressures was calculated in the model in an attempt to determine how the sinks of water vapour were organized. In the second January the maximum relative humidity (with respect to ice) is found at 200 mb. At 100 mb however, almost the entire globe has relative humidities below 100%, indicating that the desiccation of the air has been completed between 200 and 100 mb.

At 200 mb the regions of highest relative humidity were found mainly at tropical latitudes, whereas at 300 mb they lie mainly outside these latitudes. No consistent relationship between these regions and synoptic features could be determined. The correlation between water mixing ratio and temperature for all the grid points between 30ºN and 30ºS is shown for 100 mb in Fig. 23(a) and for 200 mb in 23(b). In both figures the mixing ratio corresponding to saturation with respect to ice at a certain temperature, computed using the model algorithm, is shown by the curve.

At 100 mb only a few of the points have mixing ratios greater than the saturated value (Fig. 23(a)). The mass mixing ratios cluster weakly around a value of about $3 \times 10^{-6}$ with only a small proportion approaching zero. At 200 mb (Fig. 23(b)) a larger number of points have mixing ratios greater than saturation. A weak cluster may be identified, consisting of a band of points at which there is supersaturation with respect to ice. This problem of local supersaturation occurs only at 300 and 200 mb, and is probably accounted for by the effect of the interpolation between saturation over ice below 233 K and over water at higher temperatures, combined with the formulation of condensation in the convective scheme.

There is some evidence in Fig. 23(a) that the shape of the saturation curve at a level below 100 mb is preserved in the $q, T$ scattergram; Fig. 23(b) shows conversely some evidence that at 200 mb the $q, T$ correlation at the 'half level' above is not completely lost. Relative humidities at both 200 and 100 mb in the model do not support a view of the cold trap involving uniform gentle ascent with ubiquitous condensation in the upper tropical troposphere, an interpretation consistent with satellite pictures of the distribution of cumulonimbus tops and anvils.

10. DISCUSSION AND CONCLUSIONS

A general circulation model may be used to identify modes of transport that could occur in the atmosphere and are worthy of study using atmospheric data. However, a GCM has limited spatial resolution and is unable to simulate directly processes that occur on scales less than the spacing of the grid. For example Danielsen (1982) has suggested that the desiccation of air entering the stratosphere in the tropics occurs above the tops of cumulonimbus clouds by an injection of ice crystals into the stratosphere, followed by their descent back into the troposphere. To simulate this process accurately would require a model with a much higher resolution in space than is feasible at the moment, or a numerical cumulonimbus model. Instead, it is necessary to try to describe the complexities of the process using a few basic parameters. Another consequence of this
Figure 23. Model water vapour mass mixing ratio versus temperature at day 360 (January). Measurements at all grid points between 30°N and 30°S are shown as crosses; curves join the circles defining the saturation mixing ratios. (a) 100 mb.

Figure 23(b). 200 mb.
limited resolution is that the grid upon which the computations are framed is unable to represent accurately the 'breaks' in the conventionally defined tropopause near the subtropical and polar night jet streams.

Despite these limitations and the fact that the structure of the simulated planetary waves was found wanting in some respects by Keeping (1979), the GCM has simulated the observed fluxes of water vapour well in a qualitative sense, where these are available: below 400 mb. In the model most of the water enters the stratosphere near the tropopause breaks by quasi-horizontal advection from the upper tropical troposphere into the lower extra-tropical stratosphere, although an appreciable amount moves vertically in the same region and at rather higher latitudes. This advection occurs chiefly through the action of eddies. No observations of this motion as a large-scale and systematic phenomenon have yet been made, but Smith (1968), Cluley and Oliver (1979) and Shapiro et al. (1980) have shown evidence of air, with characteristics indicative of tropospheric origin, above the tropopause. Dobson (1973) suggested that the minima often observed at about 15 km in ozone profiles could originate from the horizontal transport of tropospheric air at tropopause breaks near subtropical jet streams.

The simulated transport of water vapour by eddies in the stratosphere is greater in the northern than in the southern hemisphere, possible because direct or indirect effects due to topography are greater in the north. However, near the 'breaks' in the model tropopause there is little difference between the hemispheres in the meridional transport by eddies although the transport is greater during winter than during summer in both cases.

Longitudinal variations in the distribution of temperature in the tropics have been shown to exist both in the model and in the atmosphere. In the troposphere low temperatures correspond to areas of ascending air and hence high mixing ratios of water. However, this air will become drier via the effect of the low temperatures on vapour pressure. The effect of the localized ascent will be to cause the temperature to be lower than the zonally averaged temperature at the tropopause. The mixing ratio of the air in the lower stratosphere will therefore tend to be lower than that suggested by the zonally averaged temperature at the tropopause, a finding which does not agree with a cold trap mechanism requiring ubiquitous gentle ascent in the upper tropical troposphere.

Angell (1981) has shown that the rainfall due to the Asian monsoon, the temperature of the tropical tropopause and the sea surface temperature in the eastern Pacific Ocean are related to the phase of the Southern Oscillation and hence to the strength of the Hadley circulation. From the work of both Lorenc (1982) and Johnson and Townsend (1981) it would seem likely that the amount of air ascending over the Indonesia region (and the amount of water entering the stratosphere in this region) are also related to the monsoon. Angell also compared the water vapour measured at Washington D.C. by Mastenbrook (1974) with the temperature at the equator at 100 mb. The variations in tropical temperature at 100 mb were found to be governed by the quasi-biennial oscillation and were not related consistently with the extra-tropical mixing ratio of water vapour at Washington.

The validity of the circulation proposed by Brewer (1949) to explain the distribution of water vapour in the atmosphere has been discussed by Robinson (1980). Newell and Gould-Stewart (1981) have suggested that this circulation is manifested in the horizontal plane by ascent in certain preferred regions, notably the west Pacific near Indonesia. It seems that the GCM simulates this ascent during the northern winter, and it also produces a positive correlation between tropical lower stratospheric water vapour and ozone when both have low values, as may be seen in Figs. 1(a) and 5(c) of Allam et al. (1981), whose
Fig. 1(b) indicates that the same correlation is present in the atmosphere. Because there is no measured global distribution of stratospheric water vapour available as yet, it is not possible to test the model's synthetic patterns in the lower stratosphere by direct comparisons of global contour maps. However, the time series of balloon-borne frost point hygrometer ascents by Mastenbrook at Washington D.C. (Mastenbrook and Daniels 1980) and by Murumatsu (1981) at Sapporo, Tateno and Kagoshima provide, via the sampling effect of differential advection over the atmospheric column up to 30 km during an annual cycle, a quite severe test of a GCM simulation. Although the corresponding model water vapour time–height profiles are quantitatively not in detailed accord with the observations, encouraging similarities in the patterns do exist. Given the model's inability to represent explicitly cumulonimbus towers and their microphysics, and the limited resolution of the tropopause, this is probably the best that could be expected with current capabilities.

A testable mechanism offered by the model is the flux of air and water vapour from the upper tropical troposphere at the subtropical jet stream location into the lower mid-latitude stratosphere; the model's meridional gradient of frost point isopleths (Fig. 24) is also a testable and informative prediction, particularly with regard to the angle they make with the isentropes.

![Diagram](image_url)

Figure 24. Model section at 140°E of isopleths of frost point (heavy lines), isotachs (continuous lines, westerly; dashed, easterly), and isentropes (light lines). Day 360 (January 1st).

In part II of this paper the time continuity of tracers along three-dimensional trajectories is considered, to provide a complementary perspective to the fluxes computed here.
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