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Possible meridional circulations in the stratosphere and mesosphere

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SUMMARY

A calculation has been made of the meridional circulation sufficient to transport heat between the radiational sources and sinks in the stratosphere (15-55 km) and mesosphere (55-80 km). Assuming that the principal heat sources are in the region of the equatorial tropopause and in the mesosphere over the summer pole, with the corresponding heat sinks near the polar tropopause and over the winter pole in the mesosphere, the main circulation patterns are found. Below 30 km there is rising air over the equator with outflow towards both poles and descent towards the tropopause in high latitudes in both hemispheres. At higher levels there is ascent over the summer pole and descent over the winter pole with a well defined flow towards the winter pole above 50 km and a more indeterminate flow pattern between 30 and 50 km. The order of magnitude of the speed of these circulations is m sec^{-1} horizontally and cm sec^{-1} vertically.

The calculations were made for the solstices, equinoxes and mid-times between them and several particle trajectories were determined. The results are in agreement with many of the observed physical features of the stratosphere and mesosphere, particularly the distribution of tracer elements.

1. INTRODUCTION

During recent years, studies of the balance of energy, angular momentum and water of the atmospheric circulation in the troposphere have tended to emphasize the importance of transport by horizontal eddies compared with that by purely meridional circulations. It has also been established that, during the polar night at least, horizontal eddies are an extremely important feature of the circulation up to at least 30 km. Nevertheless, the presence of a direct Hadley cell in the tropics and subtropics can hardly be questioned while, at the present time, the transport of tracers such as water vapour, ozone and atomic debris in the stratosphere is most generally discussed in terms of meridional circulations (e.g., Ramanathan and Kulkarni 1960; Machta 1960; Brewer 1949). Inferences drawn from observations of tracer distributions are, however, largely qualitative and a first attempt has been made to specify more definitely the possible magnitudes of meridional circulations, both in the stratosphere and in the mesosphere. Vertical and meridional motions sufficient to transport the heat between radiative sources and sinks as given by Murgatroyd and Goody (1958) and Ohring (1958) have been calculated. The model used is a meridional cross-section only and it is assumed that all the heat is transported by motion in meridional planes, the horizontal and vertical velocities being calculated from the equations of heat transport and mass continuity. Possible contributions to heat transport of horizontal eddies are neglected as they cannot yet be assessed with the observational data available. There is some partial justification for this as much of the region considered is very stable and the results when compared with tracer observations suggest that this assumption is not very serious. Nevertheless, their neglect leads to considerable difficulties when angular momentum balance is considered. The sphericity of the earth is taken into account in the form of the continuity equation used. Molecular conduction is also neglected, which appears to be a reasonable assumption from the work of Davies (1959). The other major assumption made in this model is that the meridional velocity is zero at both poles.

2. The equations of heat and continuity

The total rate of change of temperature dT/dt of a moving particle can be expressed as

$$\frac{dT}{dt} = \left(\frac{\partial T}{\partial t} + \mathbf{V}.\nabla_p T + \omega \frac{\partial T}{\partial p}\right) = \frac{R}{C_p}\frac{T}{p}\omega + \left(\frac{dT}{dt}\right)_R \quad . \quad . \quad (1)$$

where T is temperature, t time, $\mathbf{V}$ horizontal wind vector, $\nabla_p T$ temperature gradient in a constant pressure surface p, ω vertical velocity in pressure coordinates, R gas constant, C_p specific heat of air at constant pressure, and $(dT/dt)_R$ rate of radiative heating. The latter is the resultant sum of the solar and terrestrial heating and is usually expressed in degrees per day.

From Eq. (1), with zonal advection zero on a meridional cross-section it readily follows that

$$\left(\frac{dT}{dt}\right)_R = \frac{\partial T}{\partial t} + v\left(\frac{\partial T}{\partial y}\right)_p + \omega ST \quad . \quad . \quad . \quad (2)$$

where v is the velocity along the meridional direction y and S $(= 1/\theta\ \partial\theta/\partial p)$, the static stability, where θ is the potential temperature. Since eddy transports are neglected, each term on the right-hand side is a mean value round a latitude circle at a given pressure.

The continuity equation, allowing for the sphericity of the earth and also applied similarly to a mean meridional cross-section is

$$\frac{\partial v}{\partial \phi} - v \tan\phi = -E\frac{\partial\omega}{\partial p} \quad . \quad . \quad . \quad . \quad (3)$$

where ϕ is latitude and E the distance from the centre of the earth to the point considered.

Integrating between two latitudes ϕ_0 and ϕ

$$v\cos\phi - v_0\cos\phi_0 = -\int_{\phi_0}^{\phi} E\cos\phi\frac{\partial\omega}{\partial p}\,d\phi \quad . \quad . \quad . \quad (4)$$

In the model used, (i.e., a meridional plane through both hemispheres extending from pole to pole), it is necessary to postulate that v tends to zero near both poles at all levels, i.e. as $\phi \to \pi/2$ and $\phi_0 \to -\pi/2$

$$\int_{\phi_0}^{\phi} E\cos\phi\frac{\partial\omega}{\partial p}\,d\phi \to 0 \quad . \quad . \quad . \quad . \quad (5)$$

and that this integral is zero when $\phi = \pi/2$ and $\phi_0 = -\pi/2$ (see Section 4).

3. The data

Murgatroyd and Goody (1958) have presented estimates of the sources and sinks of radiative energy, in degrees per day, between 30 and 100 km at the solstices, by calculating the differences between direct heating by absorption of solar radiation by ozone and molecular oxygen and the long-wave cooling by carbon dioxide and ozone. Their results have uncertainties at the upper boundary due to the failure of Kirchhoff's law in this region, and around 30 km due to the neglect of line overlap in the calculations of absorption by carbon dioxide and also to the neglect of a possible contribution due to water vapour. Their results were therefore used between 35 and 80 km.

Ohring (1958) has presented results of a similar type expressed as absorption of solar radiation minus emission of infra-red radiation in langleys per minute throughout the year, separately for the height ranges tropopause to 55 km and 21-55 km. These latter calculations included the effect of carbon dioxide, ozone and water vapour although the amount of the latter above the lower stratosphere is still uncertain. In order to obtain a unified set of data throughout the whole range tropopause to 80 km, the resultant heating rates of Murgatroyd and Goody (1958) for the height range 35-55 km were converted to langleys per minute and subtracted from Ohring's values to give the remaining absorption in the 21-35 km interval. Mean values of heating rates in the range tropopause to 21 km were also found by subtraction. Using these different values and extrapolating downwards from 35 km, an estimate was then made for the solstices throughout both hemispheres of the distribution of the net radiative heating at all levels from the tropopause upwards. The shapes of the curves of radiational temperature change rates against height so obtained were in reasonable agreement with those given by Ohring for selected latitudes and seasons in the range tropopause to 55 km. It is considered that, although the assumptions and methods of Murgatroyd and Goody and those of Ohring, both as regards the solar heating and infra-red cooling calculations, are very different, a reasonably consistent picture of the overall distribution of radiational sources and sinks has emerged. The principal feature of both papers, i.e., probable heat transfer from summer pole to winter pole in the upper stratosphere and mesosphere and from equator to both poles in the lower stratosphere and their approximate magnitudes have been retained.

Temperatures and pressure heights between 20 and 100 km as presented by Murgatroyd (1957), and extrapolated as necessary, were used to calculate potential temperatures, stabilities, and horizontal temperature gradients as functions of latitude and height. From the tropopause to 20 km, mean values supplied by the Upper-Air Climatology Division, Meteorological Office, were used.

As data for the solstices only were available above about 30 km, it was decided to form the data at other seasons by fitting a curve of the form $a + b \cos t$ to the above quantities. This involves the assumption that both radiational heating and temperature are mutually in phase with the sun. Some indications that this assumption as regards temperature is not greatly in error are given by the radiosonde observations to about 40 km of Arnold and Lowenthal (1959). This assumption does not take into account the sudden warming of the polar night.

4. Computational procedure

The calculations were carried out for 10-degrees latitude bands from pole to pole, i.e., at points 85°, 75° to 5°, − 5° to − 75°, − 85° at six-weekly intervals commencing at the solstice. The summer solstice in the northern hemisphere was regarded as occurring on the same cross-section as the winter solstice in the southern hemisphere and therefore four cross-sections gave conditions throughout the year. The heights used were 16, 21, 28, 34, 40, 50, 60, 70 and 80 km.

Potential temperatures were first calculated and from them stabilities; $\partial\theta/\partial p$ was formed by centred differences except at 16 km, at which level it was assumed that

$$\left(\frac{\partial\theta}{\partial p}\right)_{16} + \left(\frac{\partial\theta}{\partial p}\right)_{21} = 2\,\overline{\left(\frac{\partial\theta}{\partial p}\right)_{16,21}} = \frac{2\,(\theta_{21} - \theta_{16})}{p_{21} - p_{16}} \quad . \quad . \quad (6)$$

and at 80 km, where a similar procedure was adopted.

The general scheme of the computation was firstly to calculate approximate values of ω from Eq. (2) assuming that $v = 0$ everywhere. Next, using Eq. (4) with the boundary conditions $v = 0$ at all heights at one pole ($\phi = \pi/2$) and forming vertical gradients from the computed values of ω (in a similar manner to that described above for θ), a set of values of v was computed. These values were then fed back into Eq. (2) to obtain a second approximation to ω, the process then being repeated until successive sets of v and ω showed negligible change. In practice, three or four iterations were sufficient. In general, however, at this stage it was found that v did not tend to zero near the other pole ($\phi = - \pi/2$).

There were then two possibilities :

(a) accepting that it was not possible to obtain zero values of v near both poles, to attempt some minimizing procedure for v at high latitudes;

(b) make adjustments to the original data so that Eq. (5) was satisfied.

Attempts to follow course (a) were not satisfactory, it being found impossible to obtain values of v less than 100-1,000 cm sec^{-1} at 85° latitude (owing to division by $\cos \phi$ for values of ϕ near 90°; see Eq. (4)). Course (b) is only justifiable when any adjustments required do not result in significant changes in the original data. A failure here would probably mean that the original assumption as regards the mode of the heat transfer processes is unrealistic, or alternatively, the original data are grossly in error. A trial computation showed that it was possible to adjust the original data to satisfy Eq. (5) by relatively small changes in the radiational heating rates. The correction procedure adopted was as follows :

Let ϵ at any level be the correction required at all latitudes to the original values of $\partial\omega/\partial p$ to obtain final values $(\partial\omega/\partial p)'$.

Then
$$\int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} E \cos \phi \left(\frac{\partial \omega}{\partial p}\right)' d\phi = 0 = \int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} E \cos \phi \left(\frac{\partial \omega}{\partial p} + \epsilon\right) d\phi \quad . \quad . \quad (7)$$

The computation therefore included the intermediate step of calculating

$$\epsilon = -\frac{\int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} \cos \phi \left(\frac{\partial \omega}{\partial p}\right) d\phi}{\int_{-\frac{\pi}{2}}^{\frac{\pi}{2}} \cos \phi \, d\phi} \quad . \quad . \quad . \quad . \quad (8)$$

for each level and hence in Eq. (4) setting $v_{-\frac{\pi}{2}} = 0$

$$v \cos \phi = -\int_{-\frac{\pi}{2}}^{\phi} E \cos \phi \left(\frac{\partial \omega}{\partial p} + \epsilon\right) d\phi \quad . \quad . \quad . \quad (9)$$

This must also satisfy the condition $v \to 0$ as $\phi \to \pi/2$.

The correction $\Delta\omega$ to ω corresponding to the values of ϵ then followed from the finite-difference equation for $\partial\omega/\partial p$. The conversion from ω mb sec^{-1} to w cm sec^{-1} vertical velocity was made at this stage by writing

$$\frac{\partial \omega}{\partial p} = \frac{\partial w}{\partial z} + \frac{w}{\rho}\frac{\partial \rho}{\partial z} \quad . \quad . \quad . \quad . \quad . \quad (10)$$

Finally, the corrections to the radiation term followed directly from Eq. (2) using the corrected values for ω and known values of all the other quantities. The resulting radiational heating terms are shown in Fig. 1. This figure may be compared with the corresponding diagram of sources and sinks of radiative heat energy in Murgatroyd and Goody's (1958) paper. It will be seen that the positions of the sources and sinks and their approximate magnitudes are substantially unchanged but that the position of the zero line between heating and cooling is slightly shifted. Below 35 km where Ohring's (1958) values were used the corrections are also small.

5. Results

The results are summarized in Fig. 2 in four pairs of diagrams referring to the solstices, equinoxes, and the mid-periods between them. Each pair consists of a cross-section from pole to pole of meridional and vertical velocities between altitudes of 15 and 80 km.

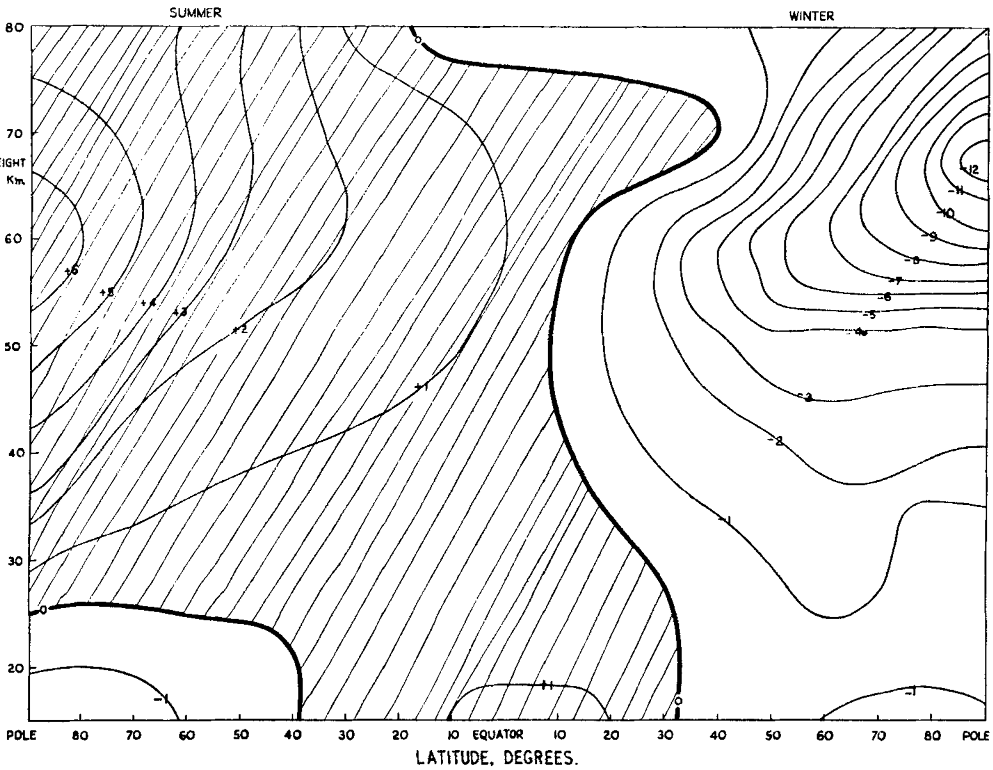

Figure 1. Radiative heating and cooling °K day^{-1}. Regions of heating are shaded.

The former have magnitudes up to a few m sec^{-1}, and the latter two or three orders less. Following the sequence of change throughout the year :

(a) In midsummer in the lower stratosphere there is a region of upwards motion near the equator of magnitude of about 0·1 cm sec^{-1} extending to about 30° latitude and, further polewards, descending air of comparable speed. Correspondingly, at all latitudes the air is moving polewards with a speed up to about 70 cm sec^{-1}. In the corresponding winter hemisphere conditions are broadly similar but the speed of the horizontal motions reaches about 100 cm sec^{-1} in middle latitudes.

In the summer hemisphere there is an equatorwards flow between 30 and 45 km with a polewards flow between 45 and 50 km both about 20 cm sec^{-1}. In the winter, apart from a small region of light equatorwards flow near the pole at 30 km, there is a general region of polewards motion of about 50 cm sec^{-1} up to 50 km. Above 50 km to 80 km, or higher, the horizontal component of motion is from summer to winter pole with speeds up to 5 or 6 m sec^{-1}. There is a continuous region of upwards motion throughout the summer hemisphere (except in the lower polar stratosphere) and downwards motion throughout most of the winter hemisphere. The maximum upwards velocity is about 1·7 cm sec^{-1} at 70 km near the summer pole, and the maximum downwards velocity 1·5 cm sec^{-1} at the same altitude and 65° latitude in the winter hemisphere.

(b) Six weeks later the general pattern has not changed except for the disappearance of the regions of polewards motion near 50 km in the summer and equatorwards motion near 30 km in the winter. The maximum speed towards the winter pole is about 4·5 m sec^{-1}. Correspondingly the maximum upwards speed over the summer pole is 0·9 cm sec^{-1} and the maximum downwards speed in the winter hemisphere is 1·3 cm sec^{-1} and somewhat nearer the pole.

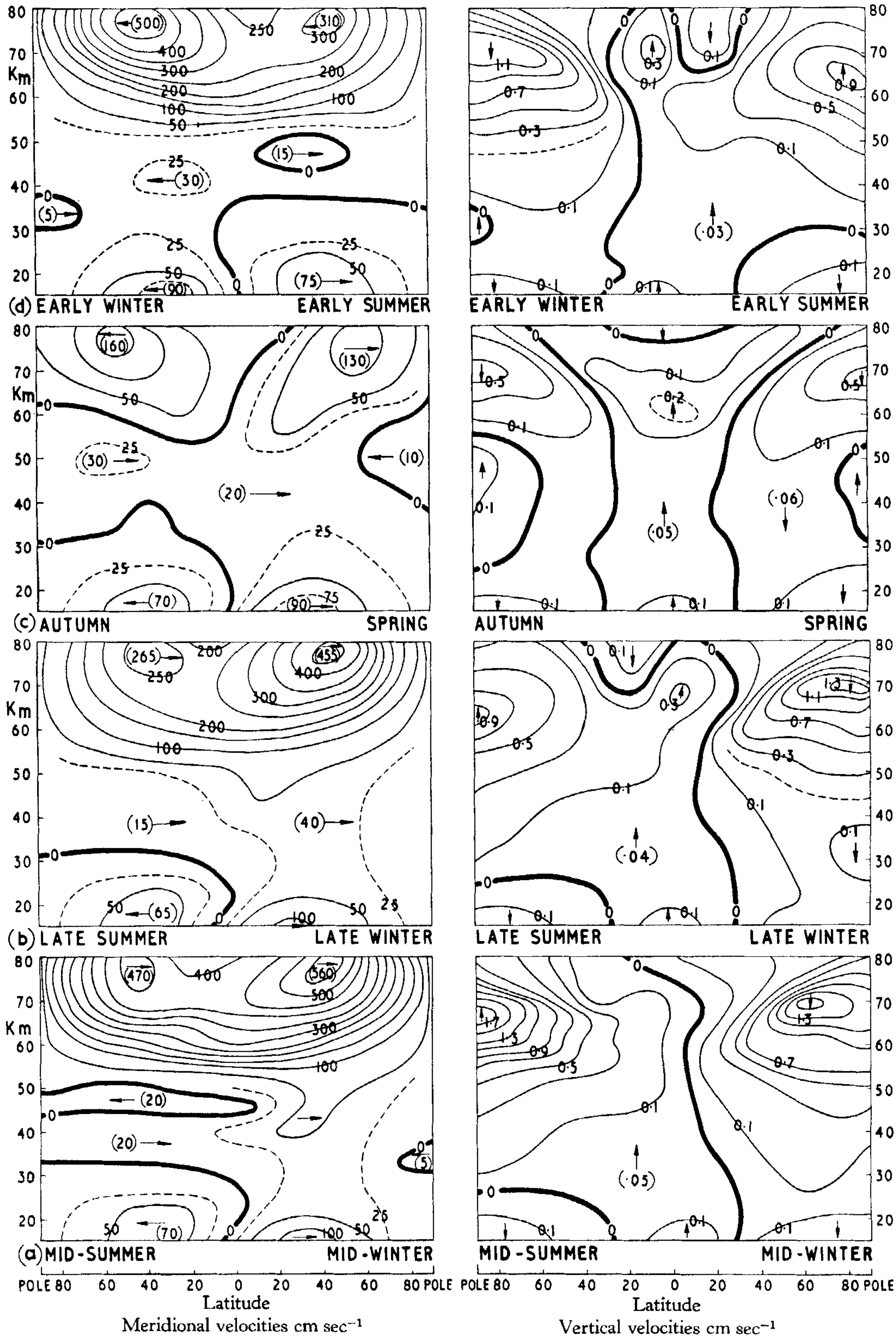

Figure 2. Cross-sections of meridional and vertical velocities (cm sec^{-1}) from pole to pole at different times of the year.

(c) At the equinoxes below 30 km there is little change, with upwards motion at the equator and downwards at both poles of about the same magnitude.

Above 30 km, however, a considerable change of pattern has taken place. The main flow above 60 km has broken up into two parts with meridional motion towards the pole in both hemispheres. A new feature is the appearance of equatorwards motion in the spring at 40 to 50 km. The horizontal motions are comparatively weak with maximum meridional speeds of 1·3 to 1·6 m sec^{-1} at 75 km.

The corresponding vertical motions are upwards in the upper stratosphere at both poles, with maxima of about 0·1 cm sec^{-1} at 45 km, and downwards, with maxima of about 0·5 cm sec^{-1}, in the mesosphere near 70 km at the poles. There is a large region of upwards motion of about 0·2 cm sec^{-1} centred at 60 km above the equator. The mechanism of change from the previous diagram appears to be the introduction of a region of ascending air in the upper stratosphere in spring and descending air near the mesopause in the autumn.

(d) Finally, as the next solstice is approached very little change is evidenced in the lower stratosphere whilst at higher levels the general pattern is tending towards the mirror image of the case (a).

It will be noted that the sequences described above are not symmetrical about the equinoxes, a result which is due to the use of the form $a + b \cos t$ to describe the seasonal variations of temperature and radiational heating.

6. Air trajectories

In order to illustrate the motion of air particles to be expected from this model a selection of trajectories has been computed and the results are shown in Fig. 3. The method was to back-track the motion in steps of three weeks, starting at selected positions in spring and autumn respectively, to examine the displacement during the preceding winter and summer half-years. In a few cases (shown dotted) the air motion was traced for a further six months. The end points were as follows :

At 20 km – 75, 60, 45 and 15 degrees of latitude
35 km – 60 and 30 degrees
50 km – 60 and 30 degrees
65 km – 60 and 30 degrees

The main features shown by this diagram are as follows :

(1) The air reaching the lower stratosphere in both spring and autumn between the equator and at least 50° latitude is always of recent tropospheric origin. Further towards the pole the air reaching the lower stratosphere in spring has descended from about 30 to 40 km during the winter. The air reaching this region during the autumn has, however, only been at heights of 20 to 28 km during the preceding summer. In both cases the continuation of the trajectories back for a further period indicates that most of the air in the lower stratosphere is of originally tropospheric origin but may have taken about 1 to 2 years to reach the lower polar stratosphere.

(2) Tropospheric air only ascends into the high atmosphere from very near the equator; descent from the high atmosphere to the lower stratosphere is confined to very high latitudes during the polar night.

(3) There is little or no transference of air between the hemispheres in the lower stratosphere.

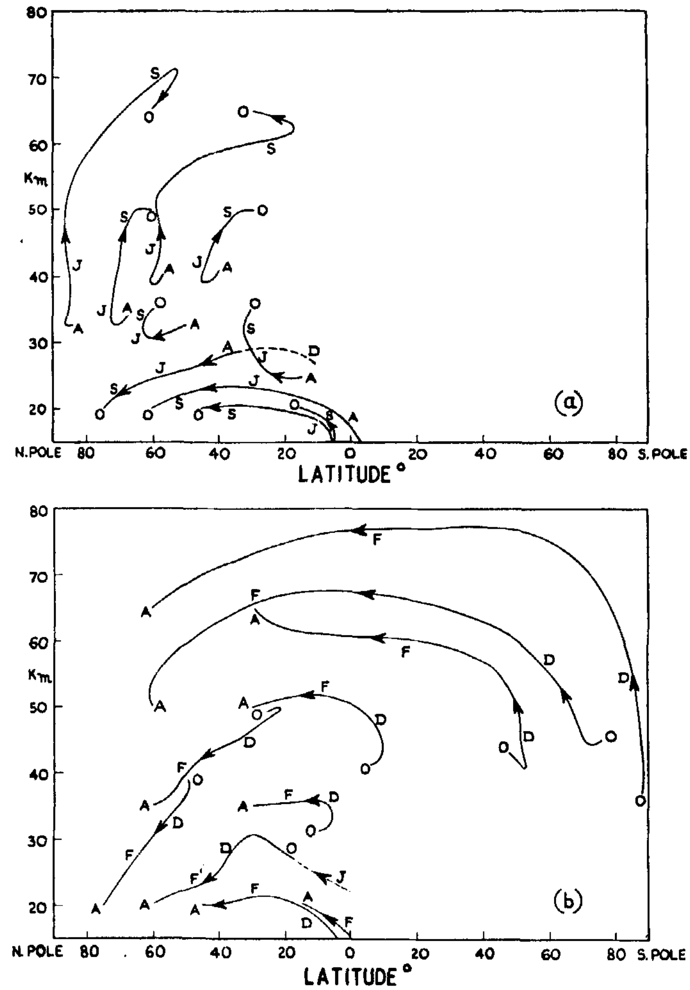

Figure 3. Selected air trajectories over six-monthly periods. Upper diagram shows trajectories ending in autumn, lower in spring in northern hemisphere. Dashed lines indicate movement in preceding six months. F denotes February position, A April, J July, S September, O October, D December.

(4) In the mesosphere, in spring, much of the air has originated in the other hemisphere and crossed the equator during the previous six months, whilst in autumn the air has generally come from lower levels in the same hemisphere.

7. Discussion

Although this model is not a very realistic representation of conditions in the atmosphere (in that it does not take into account eddy transfer processes and therefore, without further elaboration, could not be expected to be consistent with the observed mean zonal wind fields), it is, however, qualitatively in agreement with many of the other observed physical features of the stratosphere and mesophere. The idea that there was general meridional motion towards the winter pole in the mesodecline followed by subsidence in that region was put forward by Kellogg and Schilling (1951) to explain the observations of high temperatures around 65 km during the polar night. Since they could not be due to radiative effects, advection and subsidence appeared to be their main causes. Similarly, to account for the region of low temperature around 80 km over the summer pole, several writers have proposed general ascent in that region. It is possible that this could lead to the formation of noctilucent clouds if these are water (Hzostikov 1952). Alternatively, if these are not water or ice clouds but dust (Ludlam 1957), the upwards currents of order 1 cm sec^{-1} proposed above could probably support particles up to $10^{-2}\,\mu$ radius and lead to dust concentrations near 80 km in the region of the summer pole. From considerations of polarization Witt (1960) deduced particle sizes of about $10^{-1}\,\mu$. Nearer the equator, in the summer, the vertical speeds are considerably less and therefore would be less likely to support dust particles of the required size or initiate condensation. In the winter hemisphere the vertical currents are downwards and hence noctilucent clouds would not be expected.

In the lower stratosphere the scheme of circulations proposed is broadly similar to that inferred from observations of water vapour, ozone and radioactive debris. It appears from the trajectory calculations that most of the air in the stratosphere, up to 25 km at least, must have passed through the equatorial tropopause region within the last year or so. Since the latter is at a temperature of about $-80°C$ it would be expected that the frost point throughout the lower stratosphere would be around this value. This is in agreement with the many measurements by the Meteorological Research Flight in temperate latitudes; no data as yet have been obtained for the equatorial lower stratosphere. Recently there have been indications (e.g., Barclay *et al.* 1960) that the humidity mixing ratio may increase above 25 km. This is not necessarily inconsistent with the results described above as it was pointed out that there seems to be little interchange of air between the upper and lower stratosphere except very near the equator and very near the winter pole. The principal features of the observed seasonal and latitudinal distribution of ozone in the lower stratosphere are low values in low latitudes and higher values in temperate and high latitudes, where the maximum occurs during the late winter and spring. This latter observation has been shown to be due to an increase of ozone content in the lower stratosphere near the poles in these seasons. It cannot be explained by photo-chemical processes but is due rather to horizontal advection and subsidence. The trajectory calculations above suggest that these motions are largely from equator to pole with maximum heights between 30 and 40 km for air reaching the lower polar stratosphere in the spring but lower altitudes (and therefore regions of less ozone content) for air reaching the lower polar stratosphere in the autumn.

In general, air from the regions of maximum ozone content, i.e., 25-35 km tends to be transported upwards in the summer and downwards in the winter. In the former case, as it moves to regions of less ozone content as determined by the photochemical processes, ozone will tend to be destroyed. In the latter case, as the air descends the ozone content of the lower stratosphere will be increased.

The distribution of atomic debris after explosions in the lower stratosphere is such that it appears generally to move polewards, (see e.g., Hagemann *et al.* 1959; Pierson *et al.* 1960) and thereafter to have maximum concentrations in spring and minimum in autumn in the lower stratosphere. There also appears to be little transfer between hemispheres. Such motions are consistent with the circulation scheme given above with air movements tending to transport the dust upwards above aircraft sampling levels in the summer and transport it downwards again during the winter and spring.

The phenomenon of the 'sudden warming' of the stratosphere in late winter at high latitudes is not considered to be inconsistent with the ideas given above. The evidence is that (Teweles and Finger 1958) this is mainly due to large-scale subsidence and it appears likely (Stroud *et al.* 1960) that it occurs through the mesosphere as well as the stratosphere. In the large-scale meridional movements in the mesosphere suggested above, there will be continued subsidence (at a rate dependent on the radiative conditions) throughout the winter into the top of the cold polar vortex around 30 km. Whether this slow subsidence creates a general build up of warm air that finally causes the more rapid subsidence which terminates the winter regime, or whether it is some combination of this process with smaller effects linked with tropospheric conditions that finally causes the 'sudden warming,' will require further investigation. The radiative conditions and hence the rates of subsidence may be different in the two hemispheres.

It must be emphasized that these ideas are tentative and only a first attempt at providing a meridional circulation scheme for the mesosphere and stratosphere as simple as that given by Ferrel or Hadley for the troposphere. Clearly systems on the scale of cyclones and anticyclones, and possibly also, diffusion, must play a large part in the movement of the tracer elements discussed above and the model can be no more than a resultant mean effect. It is likely for instance that the filling of the lower stratosphere with ozone-rich air in the late winter will be a combination of the 'sudden warming' associated with the breakdown of large-scale cyclonic systems in conditions provided by the meridional circulations for the ozone-rich air to be available for downward transport.

This model, in company with all models which deduce meridional circulations from tracer observations, does not provide conditions for balance of angular momentum. Some of the trajectories shown, particularly at the higher levels, involve meridional displacements of tens of degrees of latitude within a few months. If angular momentum were conserved, a polewards-moving particle would be expected to acquire a large increase of westerly component, and an equatorwards-moving particle, a similar increase of easterly component. The non-conservation of angular momentum implies the existence of eddy processes which are likely to be of importance in the heat balance as well as in the momentum balance. Failure to deal fully with this point is a defect of the present model and probably the next step should be to design a further more elaborate model to attempt to resolve this difficulty. In general, the problem is more severe in the mesosphere (although when there is transference across the equator there will be cancellation effects) than in the stratosphere where the meridional velocities are considerably slower. Near the tropopause there is reasonably good agreement between the vertical and horizontal velocities and those computed from direct observations at 200 mb (Tucker 1959) and it seems that in this region at least the difficulty as regards angular momentum is not serious. Between 25 and 50 km the computed meridional velocities are in qualitative agreement with observed zonal momentum changes. It is known that between 60 and 80 km there are considerable eddy motions which probably govern the transport of angular momentum at these levels.

Finally, since the calculated circulations are supported by the tracer observations, this also provides some confirmation of the radiation data on which they are based. It is probably to be expected that the summer pole near 50 km will be a source of radiative heat due to the large direct absorption of solar radiation by the ozone layer and similarly, that at corresponding, or somewhat greater, heights in the polar night there will be a sink of energy due to lack of solar radiation. The probability of source regions near the mesopause at the summer pole and the tropopause at the equator is not so obvious. These are both very cold regions with infra-red convergence probably providing the energy source at the mesopause (Murgatroyd and Goody 1958). This suggestion seems quite plausible for the tropopause also (Brooks 1958) and has received some support by a few balloon-borne measurements (e.g., Gergen 1957). It is more difficult, however, to accept the result that the sink regions near the tropopause at the summer and winter poles are of approximately equal magnitude. Although the dispositions of the sources and sinks near the polar tropopause are probably correct the tracer observations indicate that it is likely that the winter sink is larger and of greater magnitude than that in summer.

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