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Interaction between the troposphere and stratosphere

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Three aspects of the interaction between troposphere and stratosphere are considered: (i) mass transports, (ii) radiative processes, (iii) dynamical effects.

(i) The transport of constituents across the tropopause is a major feature in determining the detailed composition of the stratosphere since the principal sources and sinks of many of its chemical species are located in the troposphere or at the Earth's surface. The nature of this transfer by the mean circulations and smaller scale phenomena is described.

(ii) Since the radiation balance throughout the atmosphere is affected by the composition at all levels, changes in radiatively important constituents in one atmospheric region may affect conditions in another. The possible importance of this mechanism in troposphere–stratosphere interaction is discussed.

(iii) It is now well established that upward transfer of mechanical energy from the troposphere by planetary-wave motions plays a major role in driving the general circulation of the stratosphere. Recent studies clarifying some details of the processes involved are outlined. The possibility that the stratospheric circulation could significantly influence tropospheric motions is also briefly considered.

Introduction

The high temperatures around 50 km in the middle atmosphere result from the strong absorption of ultraviolet solar radiation by ozone that takes place there and the general circulation in this region is, to a considerable extent, determined by its net radiative sources and sinks. It has, however, been established that the circulation of the lower stratosphere is mainly driven by that of the troposphere below and that in winter at high latitudes the dynamical effects of the latter may extend at least up into the mesosphere. As a result, the circulations of the troposphere and middle atmosphere cannot be studied comprehensively in isolation.

It is known that variability in the solar radiation, which is well marked in its far ultraviolet wavelengths and below, leads to large changes in the thermosphere. It is, however, difficult to establish with certainty either statistically or through studies of possible physical or chemical mechanisms that any effects from these perturbations are significantly transferred down through the middle atmosphere to cause observable changes in the troposphere. This leads to the belief, which is also supported from considerations of relative masses and energy contents, that lower atmosphere changes can often influence upper atmosphere conditions but not vice versa unless some so far unvalidated transfer or reflexion mechanism is found to be important.

In the discussion below of troposphere–stratosphere relations, transport of mass and constituents, radiative fluxes and dynamical interactions will be considered in turn and attention drawn to areas where further progress might be made within the Middle Atmosphere Program. Since the atmosphere must primarily be regarded as one thermodynamical system rather than a number of separate layers it is desirable first to consider the nature and characteristics of the boundary between the troposphere and the middle atmosphere – the tropopause.

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Tropopause structure

The tropopause is defined as the level at which the general tropospheric lapse rate is replaced above by an inversion (mainly at low latitudes and in summer at high latitudes), or an isothermal profile or a considerably decreased lapse rate, broadly the level at which it becomes less than 2 K/km (see W.M.O. 1966 for the detailed specification). In the last case, which occurs in high latitudes in winter, the tropopause level may be indistinct and some authors (e.g. Court 1942) suggest that the tropopause disappears at times in winter over Antarctica.

The existence of a tropopause is usually thought to be determined, at least as an annual mean on the global scale, as the demarcation level between the stratosphere where the radiation balance is the principal process determining the thermal structure and the troposphere where convective turbulent exchange is of major importance (see, for example, Goody 1949). It has been proposed (Staley 1957) that divergence of vertical enthalpy flux in the presence of the stable stratosphere above and the less stable troposphere below can lead to the formation of a sharp discontinuity in the temperature profile. Figures 1 (*a*), (*b*) illustrate the broad features of global tropopause levels and temperature distributions of the type that have to be accounted for and figures 1 (*c*), (*d*) the corresponding temperature fields in the troposphere (600 mbar)† and in the lower stratosphere (50 mbar) to which they might be related. These data were extracted from monthly mean I.G.Y. pole to pole cross sections at 75° W, July 1957–December 1958 (U.S. Department of Commerce 1961) and, although local in the sense of including some standing wave features, appear to be generally consistent with the available global climatological data. With reservations regarding their coverage and the somewhat artificial tropopause specification for the high latitudes in winter they show:

(i) Mean tropopause heights are greatest in low latitudes where the equatorial tropopause is around the 17 km or 100 mbar level. The annual variation here shows maximum heights and lowest temperatures (about −82 °C) in January and this pattern extends further into the Southern than the Northern Hemisphere.

(ii) At higher latitudes the polar tropopause at about 10 km or 300 mbar is for the most part everywhere highest in August, i.e. both in the Northern Hemisphere summer and the Southern Hemisphere winter. The temperatures are, however, highest in both their summers. Irregularities occur particularly in the Northern Hemisphere.

(iii) Comparison between figures 1 (*d*) and 1 (*b*) indicates a close correlation between 50 mbar and tropopause temperatures. On the other hand the 600 mbar values (figure 1 (*c*)) show little variation with season at low latitudes, whereas the tropopause temperatures show a well marked minimum in January. The correspondence is better at high latitudes where both figures 1 (*b*) and 1 (*c*) show warm summer and cold winter temperatures, but the detailed agreement in the more variable Northern Hemisphere is not as good as that between the 50 mbar and tropopause temperatures.

In so far as these data are representative they indicate that tropospheric convection has a major effect on tropopause height, particularly at low latitudes and in determining the overall latitude distribution. However, at higher latitudes the tropopause height and temperature appear to be strongly influenced by stratospheric temperatures tending to impose a lid or to prolong the lapse rate region upwards. The annual variations at the equatorial tropopause may be

† 1 mbar = 10^2 Pa.

linked with an annual variation of the strength of the Hadley cell (Reed & Vlcek 1969; Newell *et al.* 1969) and it also appears likely that they are a dynamical consequence of the greater activity in the Northern Hemisphere winter.

The mean tropopause is also not continuous from equator to pole but has a major gap at about 40° N in the jet stream region between the equatorial and polar tropopauses. This gap is thought to be due to dynamical effects, with widespread subsidence in this region linked to the descending branch of the Hadley cell (Manabe & Hunt 1968). Isentropic quasi-horizontal

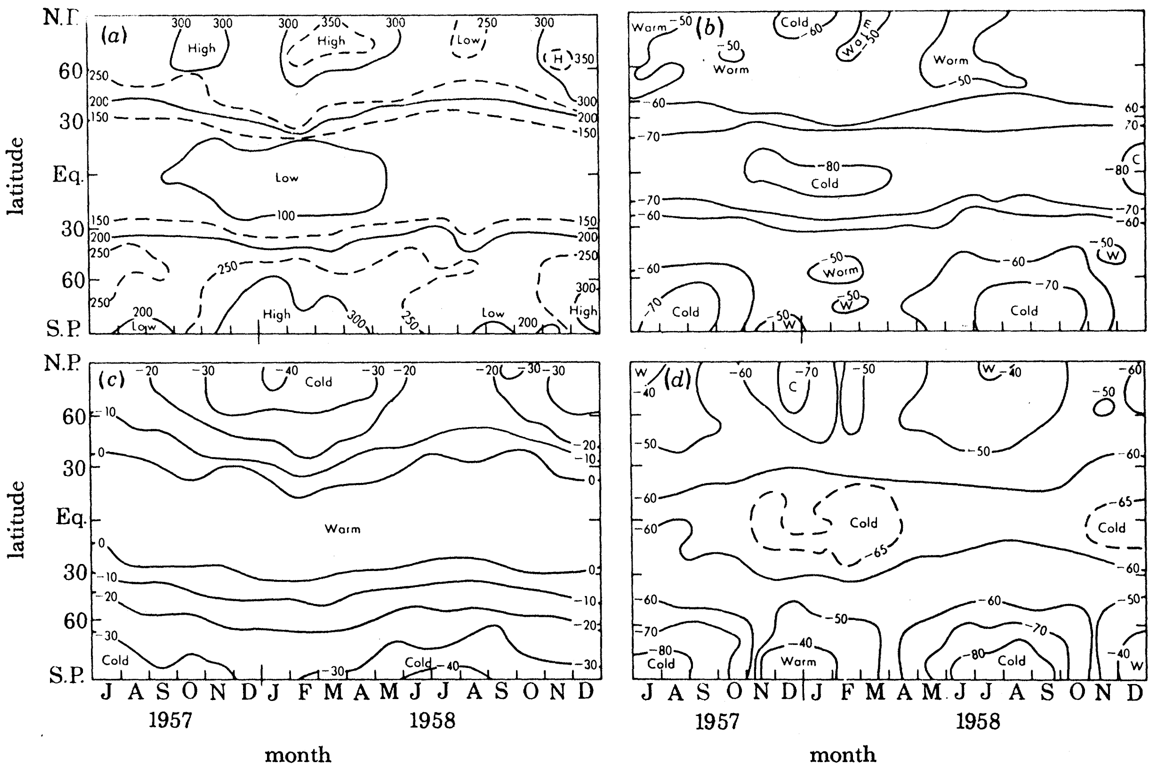

FIGURE 1. Latitude–time sections of (*a*) tropopause pressure (mbar), (*b*) tropopause temperature (°C), (*c*) temperature (°C) at the 600 mbar level, (d) temperature (°C) at the 50 mbar level. Data based on monthly mean I.G.Y. pole to pole cross sections at 75° W.

transfer of atmospheric constituents, radioactive tracers, etc. can readily take place through this gap (see, for example, Danielsen 1959). On the other hand it is not so clear that direct transfer can readily take place through an unbroken tropopause, e.g. in the upward transfer of mass and water vapour, etc. through the equatorial tropopause, although penetrative convection (see, for example, Riehl & Malkus 1958) appears to be important there, changing the problem to one of mesoscale or smaller scale transfer across the tropopause with some diabatic contributions.

There are also difficulties in explaining why the higher latitude tropopause should in general retain so efficiently its sharp change of lapse rate on the small scale in the presence of the large scale dynamical processes which take place in both troposphere and lower stratosphere.

Troposphere–stratosphere mass and constituent transfer

Reiter (1975) has attempted to construct a budget for the annual change of stratospheric mass of a hemisphere. He regarded the possibly important processes as (*i*) seasonal changes in the height of the mean tropopause level, (ii) mean meridional circulations, (iii) large scale eddy transports mainly in jet stream regions and (iv) mesoscale and smaller scale transport across the tropopause. His estimate for (i) was that the seasonal change of tropopause heights accounts

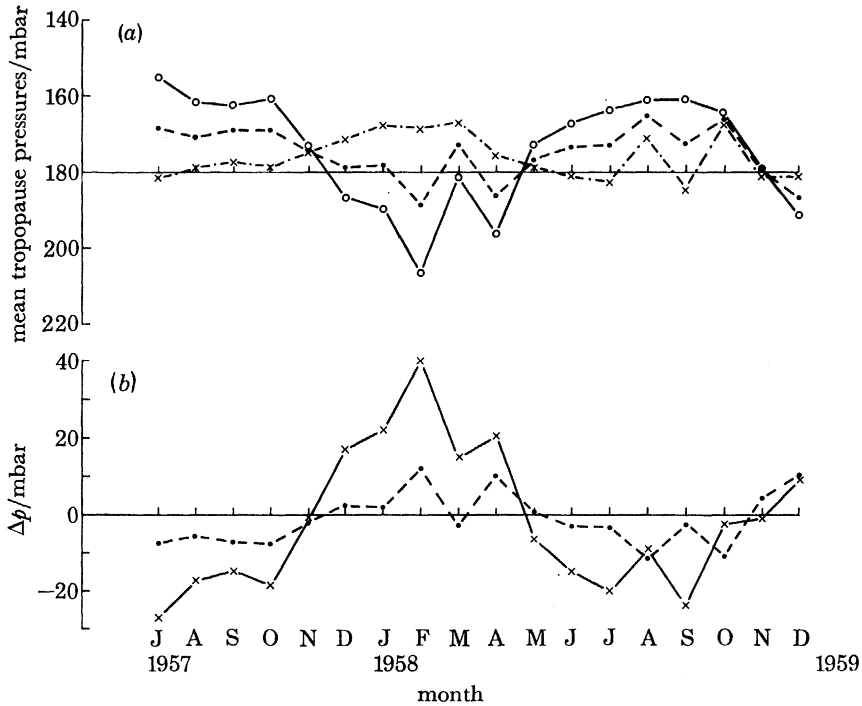

Figure 2. (a) Variation throughout the I.G.Y. of monthly mean tropopause pressure (mbar) at 75° W of the Northern Hemisphere average (○—), the Southern Hemisphere average (×–·–) and the global average (●––); (*b*) corresponding variation of Northern Hemisphere minus Southern Hemisphere values of the tropopause pressures shown in (*a*) (×—) and the deviations of the monthly global mean values from the global overall mean for the period (●––).

for a flux of about 10 % of the mass of the stratosphere in one hemisphere during the course of 1 year and he concluded that this flux is balanced approximately by the seasonal shift of stratospheric air masses between the Northern and Southern Hemispheres. Figure 2, which is based on the same data as figure 1, indicates that for the whole globe there is about 5 % more mass in the stratosphere in the November–May period than in the June–October period, the pattern being the same as that for the Northern Hemisphere but opposite to that for the Southern Hemisphere. These differences between Northern and Southern Hemispheres (figure 2(*b*)) show about 10 % more mass in the Northern than in Southern Hemisphere in the November–May period and the opposite in the remainder of the year. A further breakdown into contributions from the higher (> 30°) and lower (< 30°) latitudes indicates that the major cause of the above annual variation is that in the northern high latitudes. Conditions in the other regions show less definite variations and in the Southern Hemisphere the conditions in its summer 1958 were very different in this respect from those in 1957. Reiter also estimated that the Hadley cell is the most effective of the transport mechanisms and that it transfers approximately 38 % of the mass equivalent to one hemispheric stratosphere (which is about 40×10^{19} g) through the tropopause per year. This was based on stream function data computed by Louis (see Reiter *et al.* 1975), which are in general agreement with those of other authors. Louis also calculated that a total mass of $5–8 \times 10^{19}$ g of stratospheric air is shifted between winter and summer hemispheres. Finally, by using estimates of the mass transferences per cyclone and the cyclone frequency in the Northern Hemisphere, Reiter has estimated that systems on this scale transfer about 20 % of stratospheric air through the troposphere each year. His corresponding estimate for the smaller scale systems is only about 1 %.

The above picture of tropospheric–stratospheric mass exchange is consistent with that obtained from studies of the vertical transfer of individual constituents and their estimated stratospheric residence times. The monthly variations of total ozone over the two hemispheres and over the globe appear to be very similar although a few months later in phase than those of the tropopause heights shown in figure 2. According to London *et al.* (1978) the total ozone deduced from ground based data in the Northern Hemisphere is a maximum in March or April and a minimum in September or October. In the Southern Hemisphere the variation is in antiphase with that of the Northern Hemisphere, with the maximum in October or November. The global cycle broadly follows that of the Northern Hemisphere which has 2–3 % more total ozone than in the Southern Hemisphere. This trend is similar to that obtained from satellite results given recently by Keating (1978). Angell & Korshover (1974) have reported that the tropopause pressure and temperature exhibit a quasi-biennial oscillation. The amplitude of the pressure oscillation during the period 1957–71 was about 4 mbar at the equator and there was a pronounced out of phase relationship between equatorial and mid-latitudes. In addition these authors found that during this period there was a long term increase in tropopause pressure with values of about 5–7 mbar per decade in the tropics and 2–4 mbar per decade in higher latitudes in the Northern Hemisphere. The tropopause pressure increases were accompanied by temperature increases which may have led to increases in stratospheric humidity over the period 1964–70, observed by Mastenbrook (1971) at Washington, D.C., if it is controlled by the 'cold trap' mechanism suggested by Brewer (1949). Newell & Wu (1978) have presented evidence that latitudinal distribution of total ozone and lower stratosphere temperature vary with the intensity of the Hadley cell. They also pointed out that the long term trend in observed Northern Hemisphere total ozone amounts, which showed an increase in the 1960s, and Angell & Korshover's results above are compatible with a decrease in the Hadley cell circulation during this period. It also appeared from Newson's three dimensional model (C.O.M.E.S.A. 1975) that on the whole the flux of water vapour in winter is from stratosphere to troposphere mainly in the descending part of the Hadley cell. A mean meridional circulation with descent in middle latitudes in the winter stratosphere has also been inferred from observations of radioactive substances (List & Telegadas 1969). Descent through tropopause gaps of radioactive substances, ozone, potential vorticity and dry stratospheric air has been verified in many observational studies.

It is difficult to obtain from observations a measure of the overall rate of removal of a substance from the stratosphere and it is usually estimated in terms of its apparent residence time in the region tropopause to, say, the 25 km level. For example according to Krey *et al.* (1974), this residence time for Sr^{90} is about 14 months, which Reiter argued is consistent with a combined effect based on his estimates of vertical mass transfer by the Hadley cell and large scale eddy mechanisms together with 15 % by movement between hemispheres annually. Other authors believe the above residence time is an underestimate by a factor of two or more when applied in calculations of atmospheric composition (involving the transfer of species such as N_2O, CH_4, $CFCl_3$, CF_2Cl_2 from troposphere to stratosphere and HNO_3, HCl and O_3 from stratosphere to troposphere) in current studies of possible effects of pollutants on O_3 amounts in the stratosphere. The question is of great importance both in the calculations of the possible duration of any effects and also their magnitudes and presents one of the major uncertainties in this work, particularly as there have been many one and two dimensional studies which have attempted to represent the eddy transports in terms of empirically based flux-gradient relations. Accordingly, although there is now some confidence that the mechanisms and patterns of transfer of these substances have been identified and can be simulated reasonably well in three dimensional and, at least for the mean circulation contributions, in two-dimensional models of the atmosphere, further work is needed to obtain better quantitative estimates of the magnitudes of the different vertical transfer mechanisms.

Radiative interactions

The importance of radiation, particularly the absorption of solar ultraviolet radiation by ozone, in establishing the overall temperature distribution of the stratosphere is apparent from a comparison of its calculated temperature field in radiative equilibrium (see, for example, Manabe & Strickler 1964), with the mean observed temperature distribution (see, for example, Murgatroyd 1969). The heating rates due to this absorption by ozone are calculated to be of the order of 10 K/day near the stratopause and to fall to less than 0.1 K/day below about 20 km. The importance and role in establishing the thermal structure and driving the circulation of the different radiative processes in the lower stratosphere are more difficult to assess. Significant absorption of solar radiation also takes place here in visible and infrared wavelengths by aerosols, ozone (Chappuis bands), NO_2, molecular oxygen, water vapour and carbon dioxide with small contributions from other minor constituents (see Houghton 1963). In a similar manner to the patterns produced by the absorption by ozone at higher levels, the resultant heating rates are larger in the summer hemisphere and increase towards the pole. Hering *et al.* (1967) have estimated that absorption of back-scattered radiation in the O_3 Chappuis bands may increase heating by O_3 by about 10% around the 30 km level. The patterns of cooling and heating due to divergence and convergence of the balancing atmospheric radiation, mainly in the 15 μm CO_2, 9.6 μm O_3, and the water vapour rotational and 6.3 μm bands, show cooling increasing with height above about 30 km, but heating or small cooling above the equatorial tropopause at lower levels. Elsewhere the cooling pattern in the lower and middle stratosphere generally tends to follow the temperature distribution. In the troposphere there is a general radiative cooling with maximum rates of up to 2 K/day occurring in low latitudes. The boundary here between the net heating above, due primarily to the 9.6 μm O_3 band, and net cooling below, mainly due to the water vapour bands, is at about 15 km, i.e. somewhat below the equatorial tropopause and does not completely account for the sharp inversion there. It appears also from the calculations of Manabe & Strickler (1964) that the variation of the concentrations of the radiatively important constituents from equator to pole is not sufficient to explain the latitudinal variation of the mean tropopause height (17–10 km approximately).

The overall distribution of the net heating and cooling rates indicates that their relation to the source–sink mechanisms driving the general circulation of the troposphere and stratosphere and also the establishment of the detailed pattern of the thermal structure is indirect. Since there is net radiative cooling, which is a maximum at low latitudes in the troposphere, the contributions to the overall heating rate due to sensible heat and latent heat release must be a major factor in this region. Since the equatorial tropopause region is very cold and it is also a region of net heating, available potential energy must be destroyed in the lower stratosphere and its circulation driven by that of the troposphere. Although the heating region in the equatorial lower stratosphere is likely to favour upward motion and thereby assist the vertical extension of the rising branch of the Hadley cell, this heating primarily arises because the low-latitude lower stratosphere is a dynamically produced cold region exchanging radiative fluxes with warmer regions above and below. Accordingly in this sense at least the radiative heating region in the lower stratosphere should preferably be regarded as a by-product of the dynamics rather than vice versa.

With this background, it seems that perturbations in the infrared flux from troposphere to stratosphere are most likely to be produced by radiation in the 9.6 μm O_3 band which, as well as being the main agent in causing heating in the low-latitude stratosphere, is situated in an important atmospheric window. Consequently, in contrast to the 15 μm CO_2 and water bands which are strongly absorbent and can only 'see' the upper troposphere above about the 500 mbar level from the lower stratosphere, the 9.6 μm O_3 band may 'see' either the warm surface of the Earth through the small concentrations of tropospheric ozone or the cold tops of the clouds. According to Hitschfeld & Houghton (1961) the effects of the latter situation could be strong enough to cause all or most of the lower stratospheric heating to disappear. Ramanathan (1977) has suggested that, at least in winter, the circulation of the stratosphere can interact strongly with that of the troposphere through the agency of radiation emitted primarily by the 9.6 μm O_3 band. This downward flux is considerably enhanced during sudden warmings when both lower stratospheric temperature and ozone amounts are considerably increased. A model calculation indicated that the increase of the radiative flux into the troposphere at high latitudes during a major warming was as much as 11 % of the flux toward the poles of total energy by the general circulation and reduced the tropospheric cooling in the arctic by roughly 9 %, possibly equivalent to a temperature increase of about 2 K. Since stratospheric sudden warmings at high latitudes are accompanied by cooling at low latitudes (Fritz & Soules 1972) this mechanism could lead to a net reduction of the tropospheric available potential energy, estimated as much as 9 % in the above example. Such an interaction could lead to a significant effect on the tropospheric circulation. The situation during a sudden warming is an extreme example of this kind of radiative–dynamical feedback which is generally operative to some extent and is initiated by the dynamical forcing of the stratosphere by the tropospheric circulation, so that it appears likely that this mechanism plays a role in the overall equator to pole heat transfer by the circulation of the troposphere and lower stratosphere. Further studies are needed to provide a more detailed quantitative assessment of its mean contribution and variability.

Changes in aerosol content, ozone and water vapour in the stratosphere may affect surface heating through changing the penetration of solar radiation to the surface, altering the balance of the downward and upward fluxes of atmospheric radiation and by changing the heat balance, initiating dynamical interactions. Attempts have been made in the last few years to estimate the possible magnitude of these effects in modelling studies of differing sophistication, mostly using one- or two-dimensional convective–radiative equilibrium types, but at least one in which short integrations were made with a three-dimensional model. However, none of the present models can so far be regarded as satisfactory in their treatment of the radiative phenomena or their simulation of the atmospheric dynamics, clouds and the hydrological cycle, transfer by the oceans, etc. (see, for example, Manabe & Wetherald 1967, C.I.A.P. 1975 and C.O.M.E.S.A. 1975 for details).

For aerosols, which are of particular importance in scattering and absorbing solar radiation, there is evidence that volcanic eruptions can on occasion greatly increase the stratospheric burden. For example, it has been estimated that as a result of the 1963 Mt. Agung eruption the global mass of stratospheric aerosol increased from about 1×10^8 to 3×10^9 kg (Wilson 1970). This may have resulted in an aerosol heating rate near the equatorial tropopause of about 0.1 K/day, leading to a rise of up to 6 K in the temperature there (Newell 1970), but seemed to have had little effect on surface temperatures (Lamb 1970). In general, there appears to be a likelihood of a small decrease in the latter 1 or 2 years after a volcanic eruption (see Mass & Schneider 1977). Most of the model calculations also indicate that large increases of stratospheric aerosol amounts will cause temperature increases of perhaps around 10 K in the lower stratosphere and, probably, decreases of the order of a fraction of one degree Kelvin in mean surface temperatures. The latter would probably result also in some decrease of other forms of atmospheric energy and a decrease of rainfall. However, since it is known that in the last century the mean surface temperature has shown long term variations, whose cause has not been fully explained, of the same general magnitudes as those calculated for large increases in stratospheric aerosol amounts, it will be difficult to validate these model results completely. Considerably larger effects may occur at individual locations and further work is required to refine and expand the current results.

Several model calculations of the effect on surface temperatures of changing stratospheric ozone amounts (either by the effect of pollutants or by natural phenomena) have also been reported (see, for example, Manabe & Strickler 1964; Manabe & Wetherald 1967; C.I.A.P. 1975; C.O.M.E.S.A. 1975). It appears that a 25 % reduction might lead to about a 10 K decrease in the low latitude upper stratosphere, and a fraction of a degree Kelvin decrease in mean global temperature at the surface. It seems likely that this decrease in temperature would be enhanced somewhat if the vertical ozone profile were changed to remove more ozone from the lower stratosphere, i.e. from a typical polar ozone profile towards a typical tropical profile.

The model calculations have indicated also that increasing the water vapour in the stratosphere leads to increased infrared cooling and lower temperatures there and to very small increases in the mean surface temperatures. Estimates vary from rises in the latter of less than 0.1 K for a 10 % water vapour increase to 2 K for a fivefold increase, with the stratospheric temperature changes being about an order greater than those at the surface. If high cirrus clouds were formed, their effect would also be to cause temperature increases at the surface, perhaps about 1 K for a 25 % increase.

It seems unlikely that the large changes of stratospheric composition assumed in the model calculations could arise as the result of variability in solar radiation or other extra-terrestrial effects. Willis (1978) has shown that extra-terrestrial processes will only produce extremely small changes in water vapour in the stratosphere. Keating (1978) estimated a 3 % variation of ozone over the last solar cycle, which could be responsible for a 0.05 K change in mean surface temperature when using the C.I.A.P. (1975) results but a negligible amount if the C.O.M.E.S.A. (1975) results are correct.

Dynamical interactions

A striking feature of the stratospheric circulation is its apparent simplicity as compared with that of the troposphere. In summer, the stratospheric flow is almost completely zonal with easterly winds around a warm anticyclone approximately concentric with the pole. The winter mean circulation of the extratropical Northern Hemisphere stratosphere above about the 100 mbar level consists mainly of planetary waves of zonal wavenumbers 1 and 2 superimposed on a zonal westerly flow around a cold polar vortex, while in the Southern Hemisphere stratosphere the situation is similar except that the wave perturbations are far less pronounced. These planetary waves are quasi-stationary in phase but fluctuate in amplitude with time-scales of the order of 2 weeks. Observational and theoretical studies have shown that their associated horizontal momentum and heat fluxes play an essential role in governing the mean meridional circulation in the winter stratosphere and in inducing sometimes dramatic changes of the zonal flow, e.g. during stratospheric sudden warmings. Additionally, planetary wave motions have been found to be important in transporting various chemical species, e.g. ozone within the stratosphere. These in turn can modify the field of net radiative heating and thus the dynamical motions themselves. The predominance of planetary waves in the winter stratosphere, as compared to motions of a smaller scale, their origin and relative absence from the stratosphere in summer pose fundamental questions for dynamical theory.

There is now a considerable body of evidence (see Holton (1975) for a review) that such wave motions in the stratosphere are not excited by internal hydrodynamic instabilities of a baroclinic or barotropic nature but are produced and maintained by vertical propagation of planetary-wave disturbances generated in the troposphere. Orographic forcing and differential heating are believed to be the principal source mechanisms with the resulting wave motions being of a quasi-stationary nature. In addition, there may be some long wave generation as a result of nonlinear interactions between waves on a smaller spatial scale produced by baroclinic instability in the troposphere.

The predominance of wave motions of a planetary scale in the winter stratosphere and their relative absence from the summer stratosphere were first accounted for theoretically by Charney & Drazin (1961), their work being extended by Dickinson (1968, 1969), Matsuno (1970), Simmons (1974) and others. These studies have shown the importance of the zonal wind distribution in governing the vertical propagation of quasi-stationary waves from the troposphere. The following features have emerged: Quasi-stationary waves on a planetary scale may propagate significantly into the stratosphere during periods when the zonal wind is westerly (hence inhibited propagation in summer) and they do so such that the maximum perturbation amplitude aligns itself approximately along the axis of strongest mean westerly wind component (the so-called waveguide effect). The degree of vertical penetration of the waves depends sensitively on their zonal wavelength in that shorter waves propagate less readily, which accounts for the absence of the smaller synoptic-scale features from middle and upper stratospheric charts. A region marking the transition between westerly and easterly winds (which forms a 'critical line') acts to absorb quasi-stationary waves and to produce zonal flow accelerations in that area. Such a region exists in the equatorial stratosphere separating the westerlies of the winter hemisphere from the easterlies of the summer hemisphere. Dickinson (1968) and Matsuno (1970) have demonstrated that the resulting energy sink for the wave motions causes the upward energy flux from the troposphere to be directed mostly equatorward. Relations of the form derived by Eliassen & Palm (1960) show that, for quasi-geostrophic waves in a steady state when diabatic effects are unimportant, this implies eddy fluxes of momentum and heat directed towards the poles when the zonal winds are westerly and increasing with height, as in the winter stratosphere.

While these fluxes are of vital importance in determining the mean state of the stratospheric circulation in winter, linear perturbation analyses carried out by Charney & Drazin (1961), Boyd (1976), and Andrews & McIntyre (1976, 1978) have established that for the planetary wave disturbances to change the mean flow, wave transience, damping or interaction with a critical line are required. It has been found that transient planetary-wave activity in the troposphere is related to one of the most spectacular manifestations of troposphere–stratosphere interaction, namely the stratospheric sudden warming phenomenon for which both numerical simulations and observational studies of the real atmosphere have shown that the breakdown of the stratospheric polar-night jet, which characterizes 'major events', occurs in conjunction with enhanced upward fluxes of eddy energy from the troposphere. However, it is perhaps significant that no correspondence has yet been found between the magnitude of the energy fluxes and the intensity of the warming.

In view of the above, it is reasonable to suppose that interhemispheric differences in stratospheric circulation characteristics are related to different conditions between hemispheres in the troposphere. This is supported by detailed studies involving general circulation models of the troposphere and stratosphere (see, for example, Manabe & Mahlman 1976). In particular, numerical experiments by Kasahara *et al.* (1973) and Manabe & Terpstra (1974) have suggested that the large-scale mountain ranges in the Northern Hemisphere, such as the Tibetan Plateau and the Rocky Mountains, are chiefly responsible for maintaining the quasi-stationary waves in the troposphere. The fact that major midwinter warmings on the scale of those observed in the Northern Hemisphere have not been found in the Southern Hemisphere is therefore also likely to be ultimately related to interhemispheric differences in tropospheric forcing.

Some data relevant to the relation between the seasonal evolution of tropospheric planetary waves and those of the stratosphere have recently been obtained in the Meteorological Office from an analysis of the results of an annual integration of a three-dimensional numerical model simulating the seasonal behaviour of the troposphere and stratosphere. This model, which was based on that of Corby *et al.* (1972) and extended in the vertical by Newson to contain thirteen levels, eight of which were in the stratosphere, has been briefly described in C.O.M.E.S.A. (1975) and a fuller description of the model and its simulated climatology is in preparation.

In the examination of the seasonal variation and latitudinal distribution of stationary (monthly mean) planetary-wave amplitudes in the model atmosphere, the 5 mbar level was chosen to be representative of the middle stratosphere with the 300 and 100 mbar levels taken to illustrate variations in the upper troposphere and lower stratosphere. Figure 3 (*a*) shows that stationary wave 1, found to be the dominant mode in the stratosphere, exhibited a pronounced seasonal cycle in amplitude at the 5 mbar level with maximum amplitude in winter and minimum in summer. This is an outstanding feature of the seasonal variability of the real stratosphere and is consistent with the theoretical predictions of Charney & Drazin (1961) regarding the propagation characteristics expected for planetary waves in westerly and easterly wind regimes. However, figure 3 (*c*) shows that, in the model simulation, there was a further important seasonal variation likely to influence stratospheric wave amplitudes and thus the seasonal evolution of its circulation, namely the seasonal cycle of stationary wave 1 amplitude in the upper troposphere. This cycle was approximately in phase with that of the stratosphere. In broad agreement with observations (Labitzke & Van Loon 1972), the seasonal variation of wave 1 amplitudes in the Southern Hemisphere stratosphere was less pronounced, with amplitudes considerably smaller than those prevailing during the Northern Hemisphere winter. Figure 3 (*c*) suggests that these differences may be related to the fact that large stationary wave 1 amplitudes were sustained over a considerably longer period in the upper troposphere in the Northern Hemisphere than in the Southern Hemisphere.

A study of the corresponding fields in the model for stationary wave 2 shows that this wave also exhibited a seasonal cycle in amplitude at 5 mbar in approximate phase with that of the upper troposphere (figure 4). In the Northern Hemisphere, peak amplitudes were considerably

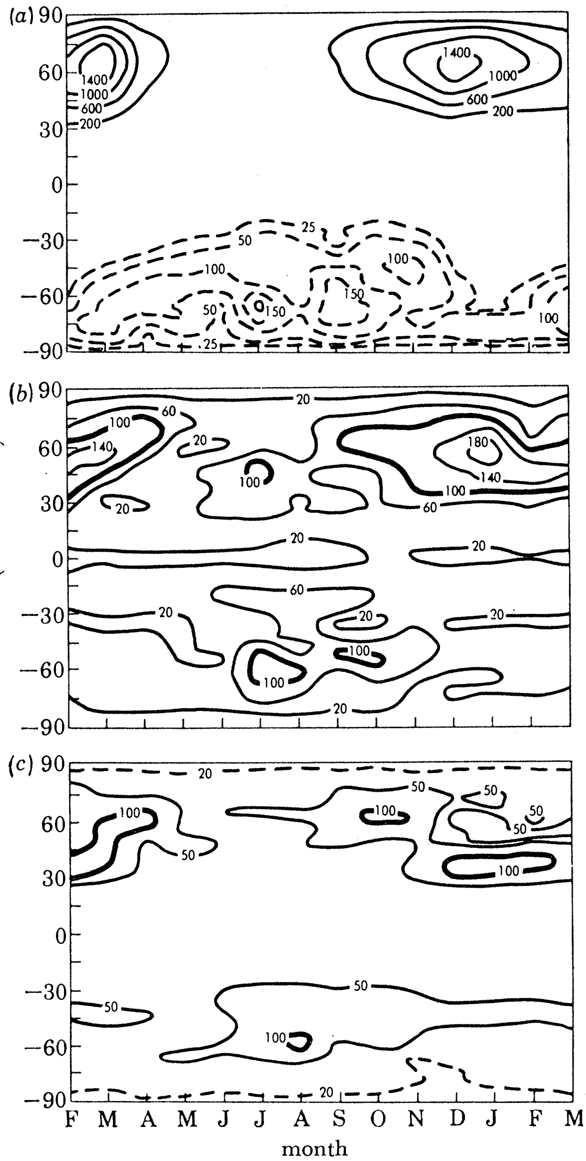

FIGURE 3. Stationary wave 1 amplitude of geopotential-height (m) at (*a*) 5 mbar, (*b*) 100 mbar, (*c*) 300 mbar.

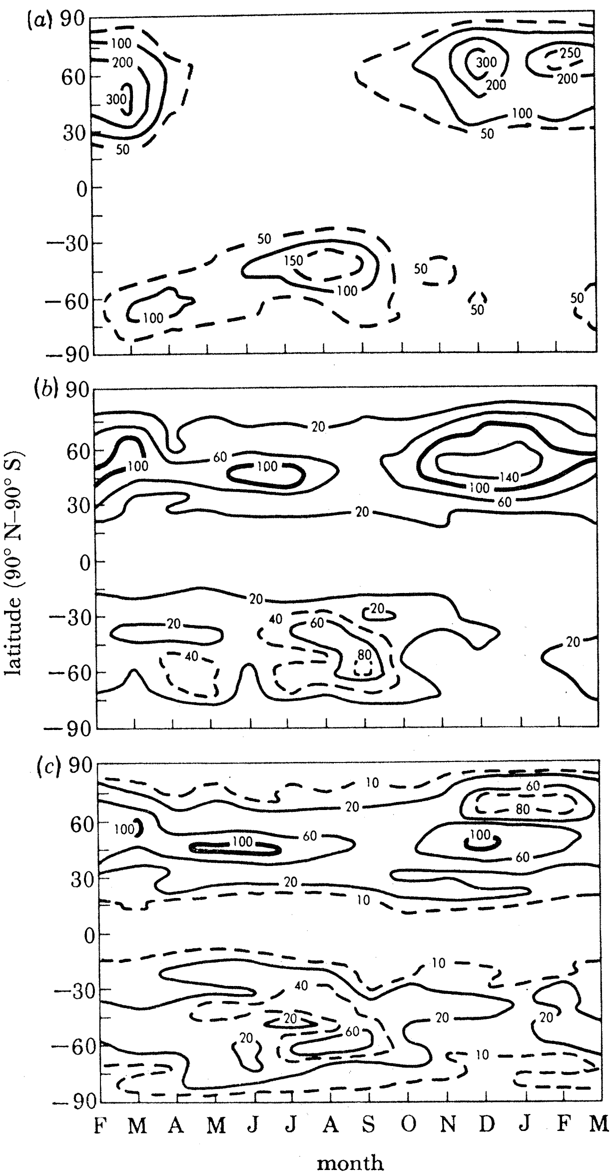

FIGURE 4. Stationary wave 2 amplitude of geopotential-height (m) at (*a*) 5 mbar, (*b*) 100 mbar, (*c*) 300 mbar.

smaller than those of wave 1 but the two waves were of roughly comparable amplitude in the Southern Hemisphere. Stationary waves of higher wavenumber than 2 were of progressively smaller amplitude in the middle stratosphere. The upward energy fluxes associated with stationary wave 1 and 2 in the model at 100 mbar are shown in figure 5, further illustrating the pronounced differences between hemispheres. Since linear wave calculations (see, for example, Matsuno 1970) indicate the preferred vertical propagation of wave 1 in the polar westerly jet, these results provide further evidence that interhemispheric differences in the stratospheric circulation, particularly in its wave 1 aspects, are directly related to those of the troposphere.

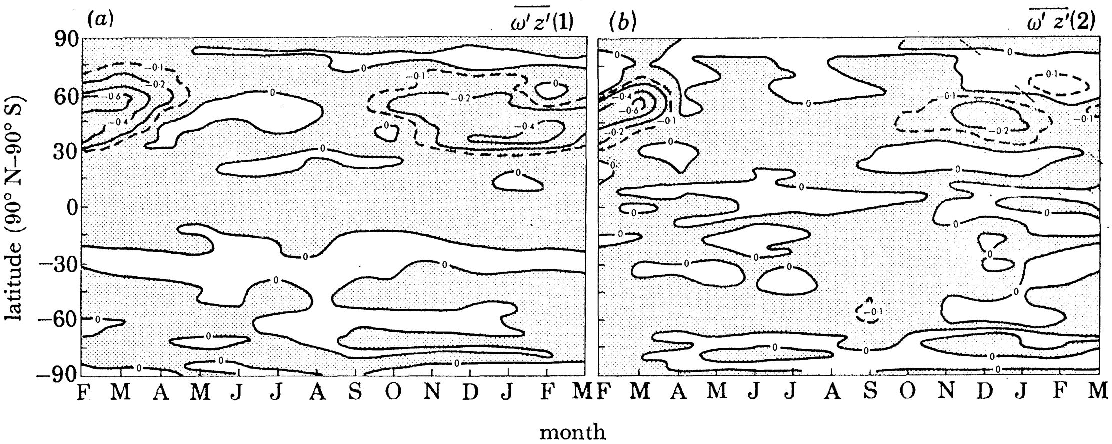

FIGURE 5. Vertical eddy flux of mechanical energy (W m^{-2}) at 100 mbar as determined by the 'pressure-interaction term' (*a*) associated with zonal wave 1, $\overline{\omega' z'}$ (1), (*b*) associated with zonal wave 2, $\overline{\omega' z'}$ (2). Stippling indicates upward flux, ω is the vertical velocity in pressure coordinate terms, z is geopotential-height, the overbar denotes a zonal average on a constant pressure surface and a prime the local deviation from that average.

Transient wave components (here taken to include perturbations with amplitude or phase differing from the monthly mean value) were of much smaller amplitude than their stationary counterparts in the Northern Hemisphere but were of comparable amplitude in the Southern Hemisphere and also exhibited a pronounced seasonal cycle in the middle stratosphere in approximate phase with that of the upper troposphere (figures 6 and 7).

In addition to the interaction between tropospheric and stratospheric circulation systems which apparently exists on a seasonal timescale, Hirota & Sato (1969) have made observations that suggest a relation between quasi-periodic oscillations in the strength of planetary waves in the winter stratosphere and the day-to-day changes in the strength of the mean zonal wind in the troposphere, the so-called index cycle. Quiroz (1969) showed oscillations of this index at 700 mbar that were in phase with the oscillations in the upward energy flux through 100 mbar, preceding thermal oscillations in the stratosphere. In a recent study of a major stratospheric warming during the 1976–77 winter in the Northern Hemisphere, O'Neill & Taylor (1979) investigated in some detail the nature of dynamical interactions between troposphere and stratosphere. They demonstrated the continuity of the planetary wave amplitudes from upper troposphere to middle stratosphere and concluded that the circulation reversal, which occurred at all levels up to at least 10 mbar, could be accounted for mainly by planetary wave 1 developments in the troposphere. In addition, they argued that Matsuno's (1971) mechanism for sudden warmings did not provide an adequate dynamical description of the event and that

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the appearance of eddy momentum fluxes towards the equator in the troposphere and stratosphere, with the consequent export of westerly momentum from high latitudes, was central to its dynamical evolution. They noted a number of other occasions during which areas of eddy momentum fluxes towards the equator in the troposphere appeared to spread upwards into the stratosphere during periods of enhanced planetary wave activity in the troposphere. Because quasi-geostrophic theory provides an adequate description of motions on this scale, it may readily be shown that this process requires that the planetary waves change their meridional structure to one in which the trough and ridge lines lie in a SE–NW direction. O'Neill & Taylor proposed that such a reorientation and amplification of planetary waves in the

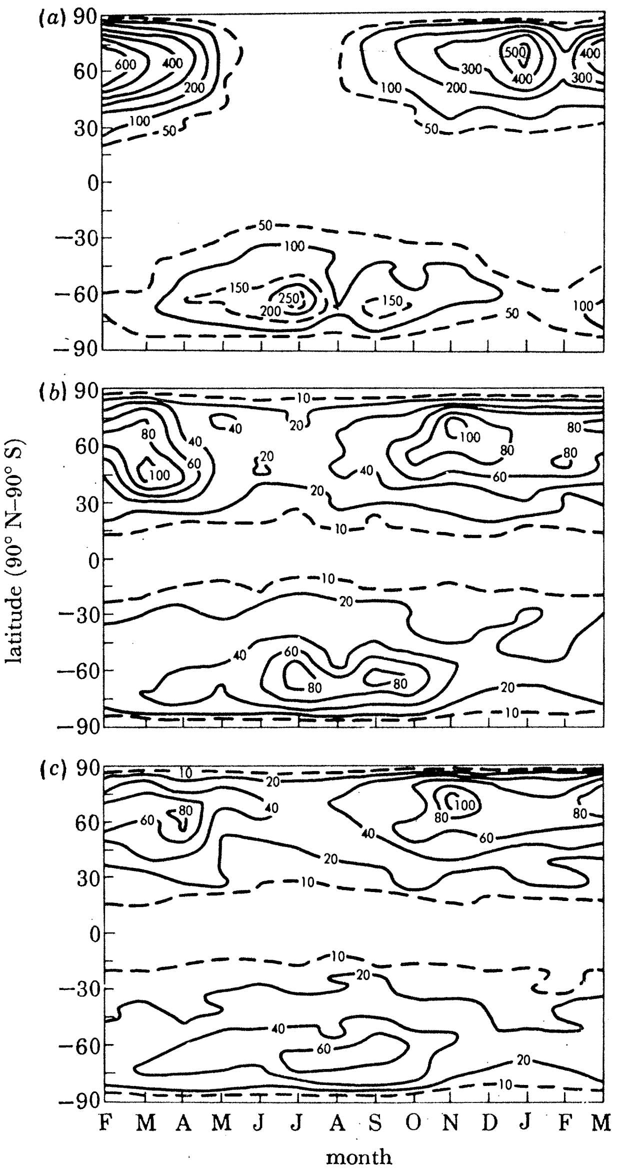

FIGURE 6. Transient wave 1 amplitude of geopotential-height (m) at (*a*) 5 mbar, (*b*) 100 mbar, (*c*) 300 mbar.

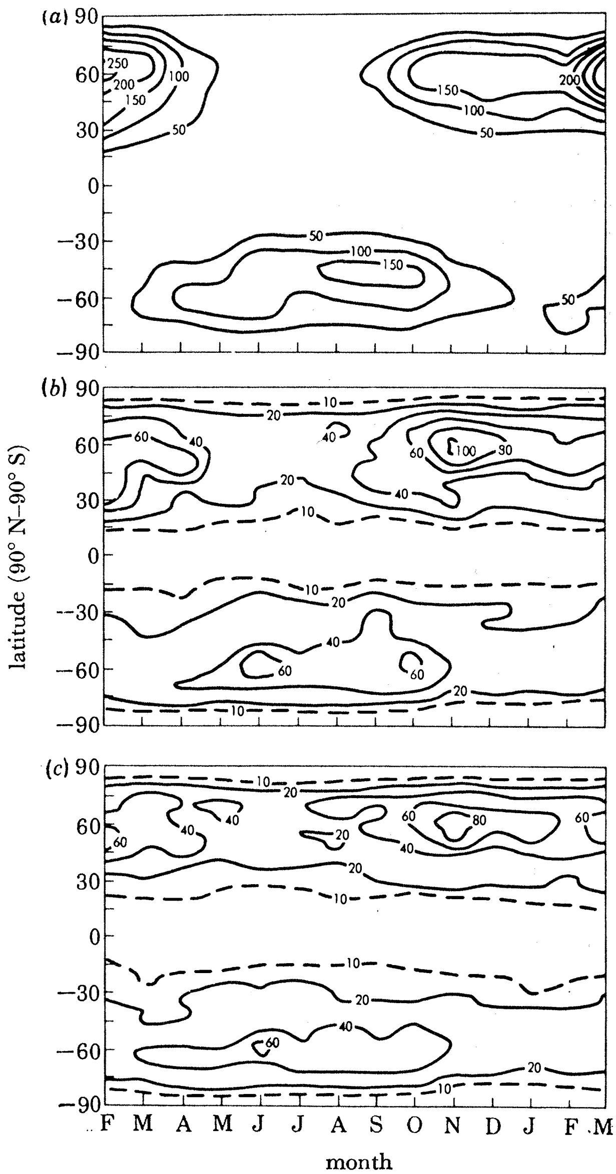

FIGURE. 7 Transient wave 2 amplitude of geopotential-height (m) at (*a*) 5 mbar, (*b*) 100 mbar, (*c*) 300 mbar.

troposphere, which they found to be associated with the appearance and subsequent evolution of a blocking-type circulation pattern, could have important consequences for the stratospheric circulation. Furthermore, both in this study and in the analysis of the above model results, it was found that major planetary wave developments in the middle stratosphere could be associated with enhanced planetary wave activity at high latitudes in the upper troposphere, low-latitude perturbations being less important in this respect. This is not unexpected in view of the location at high latitudes of the strong westerly wind waveguide formed by the polar night jet in the stratosphere. The particular tropospheric synoptic patterns associated with such developments were found by O'Neill & Taylor (1979) and to some extent in the above model simulations to include at some stage intensification of a ridge in the mid-Atlantic sector which, in conjunction with an extensive low-pressure area over east Asia, gave rise to a pronounced wave 1 component in the circulation. However, the complexity of the interactions is such that no unique relationship has yet been established between the intensity and duration of a given planetary-wave disturbance and the degree to which the zonal flow in the stratosphere is disrupted.

In the above discussion, the connection between planetary-wave motions in the troposphere and stratosphere at extratropical latitudes has been emphasized because such motions account for most of the eddy kinetic energy in the stratosphere and play an important part in determining its overall circulation. In addition, upward-propagating Kelvin and mixed Rossby gravity waves from the troposphere interacting with the wind systems of the stratosphere are believed to account for the quasi-biennial oscillation of zonal wind in the equatorial stratosphere. The former waves are also thought to act as a source of westerly momentum for the semi-annual cycle of zonal winds in the equatorial stratosphere, with planetary waves propagating upward and equatorward in the winter hemisphere providing the required oscillating source of easterly momentum. Mechanistic models which simulate the proposed interactions leading to the quasi-biennial wind oscillation have been formulated by Lindzen & Holton (1968), Holton & Lindzen (1972) and Plumb (1977).

Although a great deal of progress has been made in identifying dynamical processes by which changes in the tropospheric circulation can affect the stratosphere, there is comparatively little known of the way effects of stratospheric perturbations may be communicated downwards mechanically and perhaps cause significant alterations to the tropospheric flow. One possibly important mechanism by which this might occur is through the effect of stratospheric changes on the vertical propagation of planetary waves generated in the troposphere (Hines 1973; Bates 1977). These waves transfer their energy upwards and might be reflected at higher altitudes to interfere constructively or destructively with the initial systems that led to their generation. Hines argued that in the absence of substantial dissipation the energy density of such a reflected disturbance would be appropriate to the tropospheric level of its origin and would be likely to have some phase coherence with the generating system. It might be expected that this mechanism would be more efficient in producing a significant change in the troposphere than would be the case for an incoherent mechanism. Bates (1977), in an analytical study using a simplified model, found that the structure of its planetary-scale waves and hence their horizontal heat fluxes in the troposphere were very sensitive to his choice of stratospheric wind profile and static stability. Accordingly, he proposed that upper-atmospheric changes that might alter the planetary-wave structure could also lead to changes in the tropospheric circulation and hence an overall climatic effect. However, it was necessary for the model to contain many simplifying assumptions regarding zonal wind distribution and wave structure in order to make the problem mathematically tractable and thus more comprehensive studies are necessary before his conclusions can be generally accepted.

Discussion

Although the nature of the tropopause and its distribution in space and time are reasonably well known, there are considerable uncertainties about the precise nature of the processes by which they are determined and about the transport mechanisms in this region of the atmosphere. Until these are better understood, it will be difficult to construct the detailed budgets of the mass and constituent transports between troposphere and stratosphere necessary to gain a quantitative knowledge of the sources and sinks of the chemical species that determine the composition of the stratosphere in its 'natural' or 'perturbed' (by pollutants) state. Further studies will have to include radiative, chemical and dynamical processes on all scales.

Questions of the likely consequences in the troposphere of perturbing stratospheric composition include both effects on the penetration of solar radiation at different wavelengths and the disturbance of heat balances and hence the general circulation. These can only be answered by gaining a reasonably complete knowledge of all the physical processes that determine the natural atmosphere's composition, temperature structure and general circulation and then constructing adequate models for diagnostic and predictive purposes. It must be realized that this goal is at present far from being attained and that a great deal of work will be needed before it can be said to be reasonably approached. Work along these lines will contribute to an improved appreciation of possible solar–terrestrial relations, which can only be properly studied when the appropriate mechanisms have been quantified.

Both observational and theoretical studies have led to considerable progress in recent years in understanding the dynamical mechanisms involved in a number of stratospheric phenomena. Although it is now established that planetary-wave developments in the troposphere are important in governing the stratospheric circulation, the situation as regards possible dynamical effects of stratospheric changes on the troposphere has yet to be clarified. In the case of the quasi-biennial oscillation, it is now known that this can be observed in a number of tropospheric parameters at middle and high latitudes (Ebdon 1975; A. F. Jenkinson 1978, private communication). This raises the question of the extent to which the quasi-biennial oscillation may be intrinsic to the stratosphere and communicated to the troposphere, for example by the influence of the changing zonal wind distribution in the equatorial stratosphere on the overall planetary-wave structure.

It is hoped that in view of the theoretical and practical importance of the subjects discussed in this paper, their further investigation will form an important part of the Middle Atmosphere Program.

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