Chapter 12 Interaction between the Middle Atmosphere and the Lower Atmosphere

12.1 Introduction

The middle and lower atmospheres are linked dynamically, radiatively, and chemically: the general circulation of the middle atmosphere is to a large degree controlled dynamically by motions that originate in the troposphere; the thermal structure of the middle atmosphere is influenced by upwelling thermal radiation originating in the troposphere, and by solar radiation backscattered from the surface and clouds; and the chemical composition of the middle atmosphere is influenced by tropospheric source gases that are transported into the stratosphere and by the loss of photochemically active gases through transport into the troposphere. Thus, the troposphere strongly influences the physics and chemistry of the middle atmosphere in a variety of ways. The influence of the middle atmosphere on the troposphere, on the other hand, is far more subtle. There is, however, growing evidence from both theory and observation that the middle atmosphere does play at least a minor role in the general circulation and radiative balance of the troposphere and a crucial role in the chemistry of the troposphere.

The most obvious link between the middle and lower atmospheres is, perhaps, the direct exchange of radiation between the stratosphere and troposphere (or ground) due to emission and absorption of radiation by various trace gases. Changes in the distributions of any of these may influence the temperature profile in both the troposphere and stratosphere, and hence may possibly influence the general circulation in both regions.

The radiative linkage is complicated by the fact that many radiatively active trace gases also participate in photochemical processes that may influence the budget of ozone. For example, the synthetic chlorofluoromethanes have direct radiative effects in the troposphere due to their strong absorption in the infrared. They also may indirectly affect the temperature in the troposphere through their photochemical influence on the ozone layer. The actual role of such processes depends not only on radiation and photochemistry, but also on the manner in which transport influences the distributions of photochemically and radiatively significant tracers within both the troposphere and the stratosphere.

Tracer transport is, of course, only one manner by which the motion field provides linkages between the lower and upper atmospheres. Meteorologically more important are the links provided by the vertical propagation of wave activity as indicated, for example, by the EP flux. It is through such vertical wave propagation that the troposphere provides a first-order control on the circulation of the middle atmosphere (see Chapter 7).

In this final chapter we examine some of the radiative and dynamical processes that link the lower and upper atmospheres and discuss possible climatological consequences of such linkages.

12.2 Radiative Links: Deductions from Simple Models

The major solar and atmospheric radiative processes, which together establish the vertical temperature structure of the middle atmosphere, were discussed in Chapter 2. To a first approximation the radiative equilibrium temperature distribution is established by the balance among absorption of solar ultraviolet radiation by O_3 and the absorption and emission of thermal radiation by CO_2, O_3, and H_2O. There are, however, a number of minor trace gases that have strong absorption features in the 7- to 13-μm region where the atmosphere is otherwise rather transparent; together these may have a significant impact on the radiative balance. The concentrations of several of the radiatively active gases (including tropospheric ozone) are increasing in time due to various human impacts; the concentration of stratospheric O_3, on the other hand, is predicted to decrease (see Chapter 10). Stratospheric aerosols, which also influence the radiative balance, vary dramatically in time, mainly owing to volcanic emissions.

The possible impacts of perturbations in the concentrations of radiatively active species on atmospheric and surface temperatures have been estimated primarily with the aid of one-dimensional radiative convective equilibrium models. These solve for the vertical temperature profile in an atmospheric column by considering only the vertical transfers of heat due to radiation and convection. Normally, the temperature profiles computed are intended to represent global mean profiles. (Recall, however, that such a model was used to compute the latitudinally dependent "radiatively determined" temperature distribution shown in Fig. 1.2.) Typically, these models are solved iteratively to determine, for a given distribution of trace species, the temperature profile that satisfies the following conditions: (1) a balance between the upward infrared radiation and downward solar radiation at the top of the atmosphere, (2) radiative equilibrium at each level in the stratosphere, (3) a fixed lapse rate (usually 6.5 K km^{-1}) in the troposphere, and (4) a surface energy balance. The last may include explicit calculation of the surface-atmosphere exchange of latent heat (Ramanathan, 1981). Such models greatly simplify the actual situation by neglecting horizontal variability and many of the feedbacks among radiation, dynamics, and chemistry. Nevertheless, they are valuable tools for exploring the sensitivity of the atmosphere to changes in radiatively active species.

CO_2 is the best known example of a radiatively active gas whose concentration is changing in time. Current projections indicate that the present atmospheric abundance of CO_2 will increase by 50% during the next 50 years, due to burning of fossil fuels. Such an increase is expected to cause an increase in global mean surface temperature of about 1 K, while temperatures in the stratosphere will *decrease* by several degrees due to the increased thermal emission to space caused by the increased CO_2 concentrations in the stratosphere. The increases currently being observed in several of the minor traces gases—N_2O, CH_4, and the chlorofluoromethanes—together may produce a surface temperature increase nearly as large as that predicted for CO_2 alone (Ramanathan *et al.*, 1985). Such optically thin minor species have little direct radiative effect on the temperature of the stratosphere, but they may have an indirect influence through their influence on the photochemical budget of the ozone layer (see Chapter 10). Radiative convective models predict that increases in CO_2 should produce temperature decreases in the stratosphere that are several times the tropospheric temperature increase. An example is given in Fig. 12.1, which shows predicted vertical profiles of the atmospheric temperature change from 1980 to 2030 given by a one-dimensional radiative convective equilibrium model using a current "best estimate" scenario for changes in the concentrations of various trace gases during that period.

The large temperature decrease in the stratosphere given for the "all trace gases" case of Fig. 12.1 is primarily due to the decrease in stratospheric ozone due to chlorine chemistry. The predicted ozone change is strongly height-dependent, with maximum decreases expected near the 40-km level. Such ozone depletion should cause a substantial decrease in the solar heating due to absorption by O_3 and hence in the temperature of the upper stratosphere. Although increased ultraviolet radiation at the ground, rather than

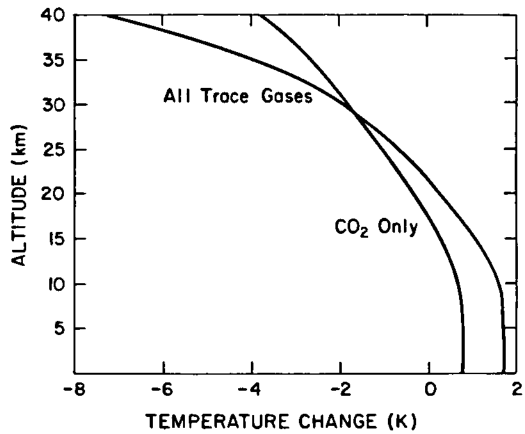


Fig. 12.1. Predicted change from 1980 to 2030 in the vertical distribution of temperature due to increase in CO_2 alone and in CO_2 along with all other trace gases that are thought to be increasing or decreasing in concentration. [After Ramanathan *et al.* (1985).]

surface-temperature change, is the primary environmental concern related to ozone depletion, radiative convective models do suggest that a decrease in the stratospheric ozone not only will decrease the temperature in the stratosphere due to the reduction of the absorption of solar radiation by ozone, but also may tend to cause a small increase in temperature at the ground in low and middle latitudes due to the increased penetration of downward-directed solar ultraviolet radiation.

At higher latitudes, surface warming due to increased solar flux at the ground is more than compensated by surface cooling due to a reduced downward flux of infrared radiation. The latter is caused by the temperature decrease in the stratosphere, which leads to smaller infrared emission for both ozone and carbon dioxide. This, then, is one example (although rather minor) of a situation in which human impacts on the stratosphere may cause changes in the troposphere.

12.3 Radiative Links: Deductions from GCMs

The prediction based on one-dimensional radiative convective models that stratospheric temperatures should be strongly affected by perturbations in the concentrations of ozone or carbon dioxide is supported by a series of three-dimensional general circulation model experiments reported by Fels *et al.* (1980). They used the GFDL SKYHI model described in Section 11.2 to examine the annual mean temperature changes that would result from a doubling of CO_2 or a halving of O_3. For these experiments the low-resolution 9°-by-10° version of the model was used. This version underestimates the eddy activity in the stratosphere and hence produces a climatology that is too close to radiative equilibrium, but it should still provide qualitatively valid information.

For both the ozone and CO_2 perturbation experiments, the zonally averaged temperature perturbations predicted in the GCM runs were compared to predictions not only from a radiative convective model applied at each latitudinal gridpoint in the GCM, but also from a two-dimensional model, called the fixed dynamical heating (FDH) model. In the FDH model a heat balance is assumed to hold at each height and latitude such that the sum of solar heating and long-wave cooling is balanced by dynamical heating or cooling:

$$\bar{Q}_{SW} + \bar{Q}_{LW}(\bar{T}) + \bar{Q}_{DYN} = 0, \qquad \bar{Q}_{DYN} \equiv -(\overline{D\theta/Dt}). \qquad (12.3.1)$$

The FDH model assumes that for any perturbation in the trace gases that alters $\bar{Q}_{SW}$, to $\bar{Q}_{SW}^{(1)}$, say, the temperature profile is altered to $\bar{T}^{(1)}$ such that a new balance arises in the form

$$\bar{Q}_{SW}^{(1)} + \bar{Q}_{LW}(\bar{T}^{(1)}) + \bar{Q}_{DYN} = 0. \qquad (12.3.2)$$

Thus, in the FDH model the zonal-mean dynamical heating at each altitude and latitude is assumed to remain fixed; the perturbed temperature is found by subtracting Eq. (12.3.1) from Eq. (12.3.2) and performing a purely radiative calculation.

The arguments presented in Chapter 7 concerning the role of atmospheric eddies in establishing the temperature structure of the middle atmosphere suggest that the fixed dynamical heating assumption is a reasonable first approximation. There it was shown that the departure of the temperature distribution from radiative equilibrium is dependent on the intensity of the eddy forcing. Thus the net radiative heating [the sum of the first two terms in Eq. (12.3.1)] should depend on dynamics and will be altered only by processes that alter the dynamics. Specifically, a change of the EP flux divergence or small-scale heating or momentum forcing terms is required to support a change in the departure of temperature from radiative equilibrium [see Eq. (7.2.4c)]. Now, most of the eddy activity in the stratosphere is believed to result from vertical propagation of eddies generated in the troposphere. Thus, perturbations in radiatively active trace gases that do not influence the climate of the troposphere should not change the intensity of the eddy sources for the stratosphere. As the eddies propagate into the stratosphere, their propagation characteristics depend on the zonal mean

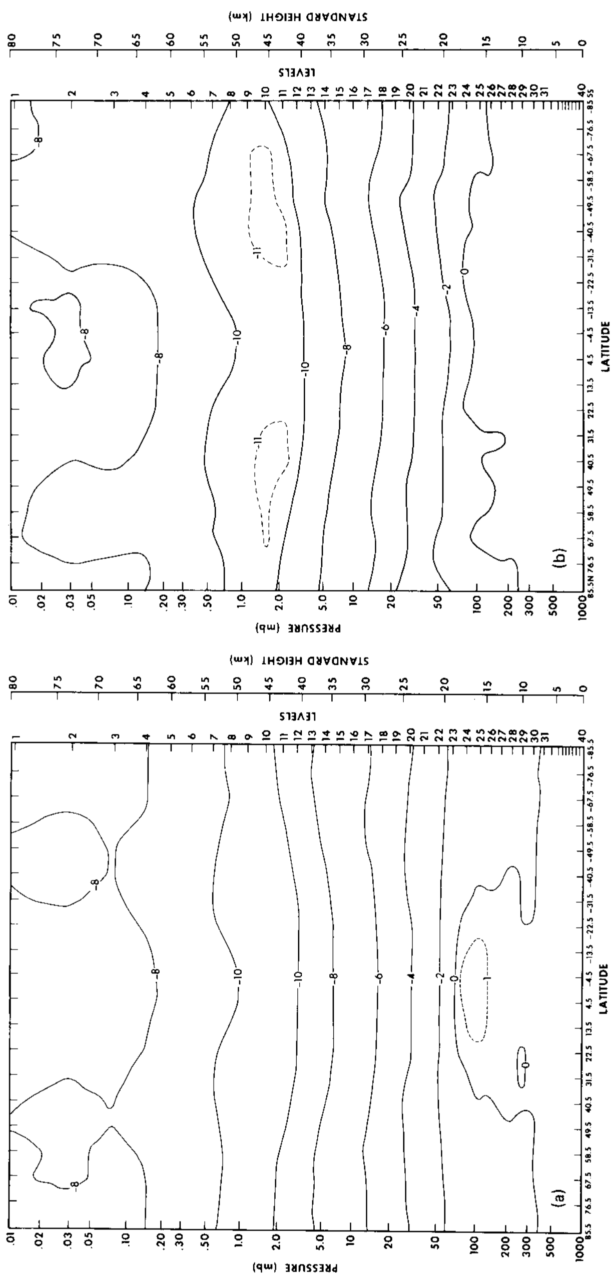


Fig. 12.2. Comparison of temperature changes simulated by (a) the FDH model and (b) a GCM for doubled CO_2 and annual mean forcing. [After Fels *et al.* (1980). American Meteorological Society.]

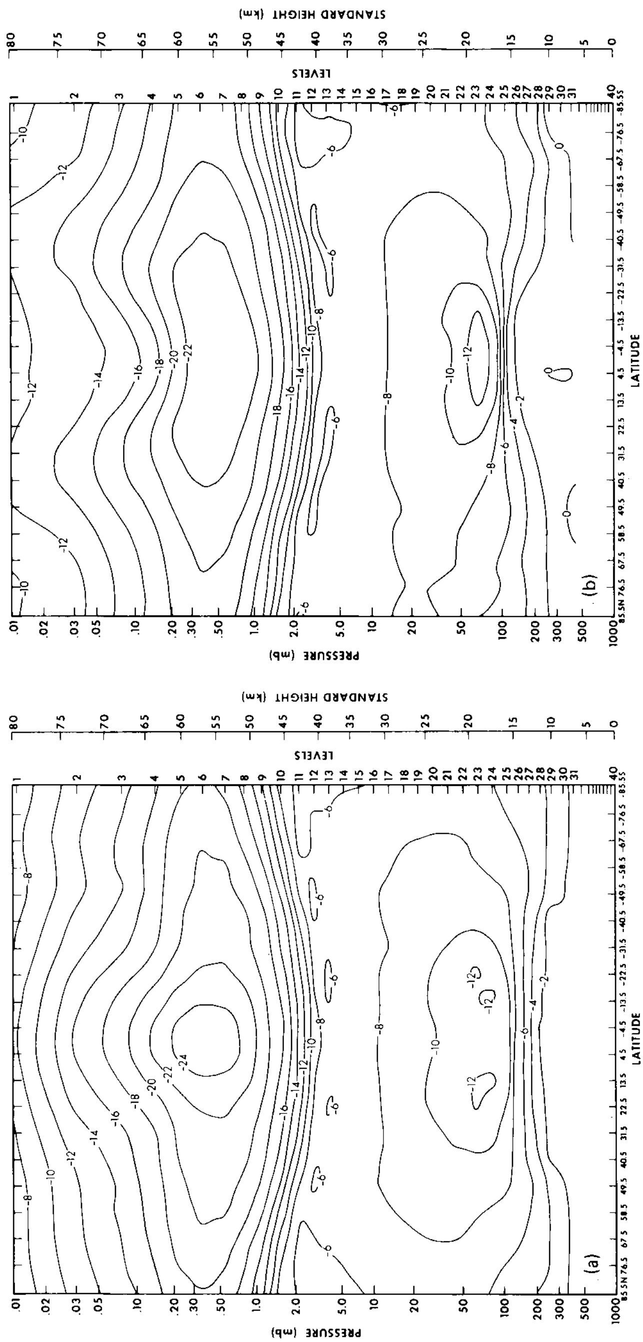


Fig. 12.3. Same as Fig. 12.2 but for 50% ozone reduction. [After Fels *et al.* (1980). American Meteorological Society.]

wind distribution. Hence, a latitudinally dependent alteration in the radiative equilibrium temperature structure will (through the thermal wind relationship) tend to alter the mean wind and hence change the eddy forcing. For temperature perturbations that have very small latitudinal gradients this effect is small, and hence the FDH assumption may be quite good.

Examples comparing the temperature perturbations computed in the GCM with those of the FDH model for the doubled CO_2 case are shown in Fig. 12.2. The same radiative heating code was used in these two calculations, and the dynamical heating produced by the GCM was used in the FDH model. Note that a temperature decrease of about 10 K, almost uniform with latitude, occurs near the 50-km level. Such a large negative temperature perturbation will have a significant effect on the temperature-dependent reaction-rate coefficients in the ozone photochemistry. The result will be an increase of ozone, and hence an increase in the ozone heating, which compensates by about 10% for the cooling due to the increase in CO_2. Because of the lack of latitudinal gradients in the perturbation, the zonal wind structures in the control and perturbed cases are very similar, and the FDH assumption is a good approximation.

The temperature perturbation for a halving of the ozone concentration is shown in Fig. 12.3. Again, the FDH model provides quite a reasonable approximation to the GCM results, even though the latitudinal temperature gradient is somewhat weaker in the FDH model so that the mean zonal winds are weaker and the fixed dynamical heating assumption is not as good an approximation as in the CO_2 perturbation case.

12.4 Dynamical Links: Vertically Propagating Planetary Waves

In Chapters 4 and 5 we discussed the influence of the zonal wind distribution in the stratosphere on the propagation and mean-flow interactions of the planetary waves that are generated by orography and heat sources in the troposphere. The observed distribution of planetary waves at any level in the stratosphere depends both on the source distribution in the troposphere and on the propagation characteristics of the atmosphere between the level of observation and the source region. Changes in planetary wave amplitudes and distributions in the troposphere due to anomalies in the source strength, or in the mean zonal wind distribution, clearly may have substantial impacts on the circulation of the stratosphere. This fact is illustrated by the discussion of stratospheric warmings in Chapter 6. It is much more difficult, however, to demonstrate that the mean wind or planetary waves in the stratosphere play a significant role in the climate of the troposphere.

In theory, the stratosphere might have a significant dynamical influence on the troposphere even if planetary wave sources were limited entirely to the troposphere, provided that some reflection of vertically propagating wave activity were to occur. An ozone reduction that was primarily limited to high latitudes, as predicted in some two-dimensional models, would alter the latitudinal gradient of the radiative equilibrium temperature, and hence should perturb the mean zonal wind distribution. This might in turn change the amount or distribution of wave reflection. However, the quasi-geostrophic model of Schoeberl and Strobel (1978) showed only a small planetary-wave response for even a large high-latitude ozone reduction.

Only if wave reflection were to create a resonant amplification of tropospheric waves does it seem possible that mean wind changes in the stratosphere could influence the climate of the troposphere. Tung and Lindzen (1979) argued that such a resonant amplification could occur if for some reason the stratospheric polar night jet were to weaken and at the same time descend to a lower than normal elevation, as observed during the initial phase of some stratospheric sudden warmings. They proposed that this linear resonance could account for planetary-scale "blocking" events in the troposphere. Plumb (1981) (see Section 6.3.4) suggested a nonlinear feedback mechanism in which amplifying planetary waves alter the mean wind distribution in the stratosphere so that it becomes closer to a resonant state, causing waves to grow more rapidly.

Resonance is a mechanism through which in principle the stratosphere might strongly influence the planetary wave distribution throughout the depth of the atmosphere, including the troposphere, and hence play a significant role in the production of climate anomalies. Thus, resonance might seem to provide a qualitative explanation for the apparent relationship between the occurrence of sudden stratospheric warmings and tropospheric blocking. As yet, however, there has been no quantitative demonstration that such stratosphere–troposphere resonances can occur in the real atmosphere. Indeed, detailed calculations of the linear stationary planetary wave response to realistic topography and heating distributions argue against the resonance theory. Using a high-resolution global primitive-equation model, Jacqmin and Lindzen (1985) demonstrated that the wave response in the troposphere is quite insensitive to changes in the basic state in the stratosphere, provided that the zonal-mean wind variation in the troposphere is limited to the range of values observed. Even large alterations in the distribution of the polar night jet in the stratosphere caused only very small changes in the stationary circulation of the troposphere in their model, although large changes occurred in the planetary-wave distribution in the stratosphere. Little evidence of planetary wave reflection or resonance was found in the several cases that they studied.

Boville (1984), on the other hand, showed on the basis of experiments with a general circulation model that unrealistically large changes in the

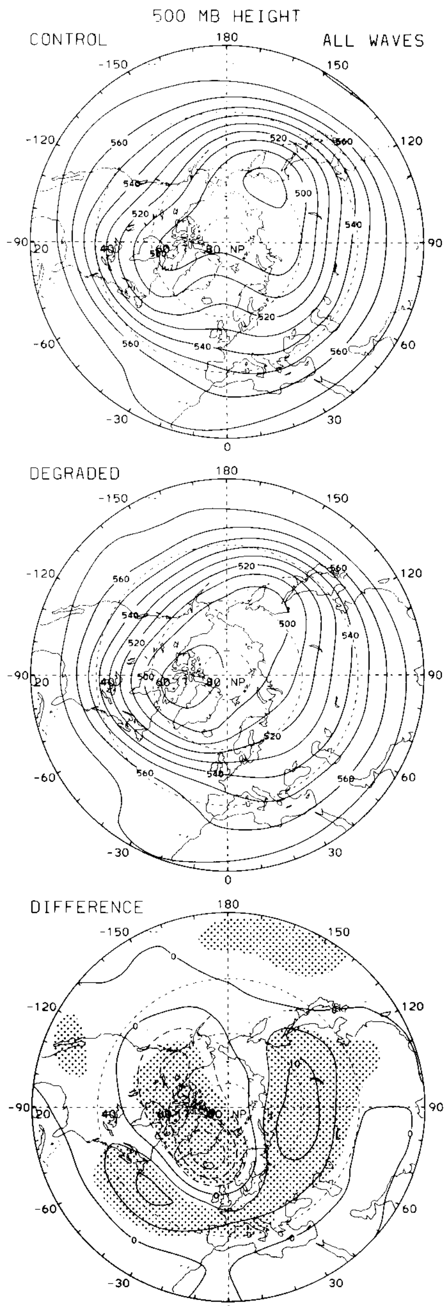


Fig. 12.4. Average 500-mb geopotential height fields (labelled in decameters) for (top) a standard January simulation with the NCAR spectral GCM, (middle) a simulation with a degraded climatology in the lower stratosphere, and (bottom) the degraded minus control difference. [After Boville (1984). American Meteorological Society.]

mean zonal wind distribution of the upper troposphere and lower stratosphere could produce dramatic alterations in the midtropospheric wave fields. These changes affected not only the stationary planetary wave structure, as shown in Fig. 12.4, but the transient cyclone-scale waves as well. The cyclone waves were apparently altered because the "storm tracks" along which they develop and evolve are strongly influenced by the stationary planetary-wave fields. Boville's results indicate that even if in reality the stratospheric circulation does not significantly affect the troposphere, it is very important to properly model the lower stratosphere in simulations of the general circulation, and in particular to avoid the common model defect of the cold winter pole bias (see Section 11.2.1) in order to avoid adverse impacts on the simulated tropospheric circulation.

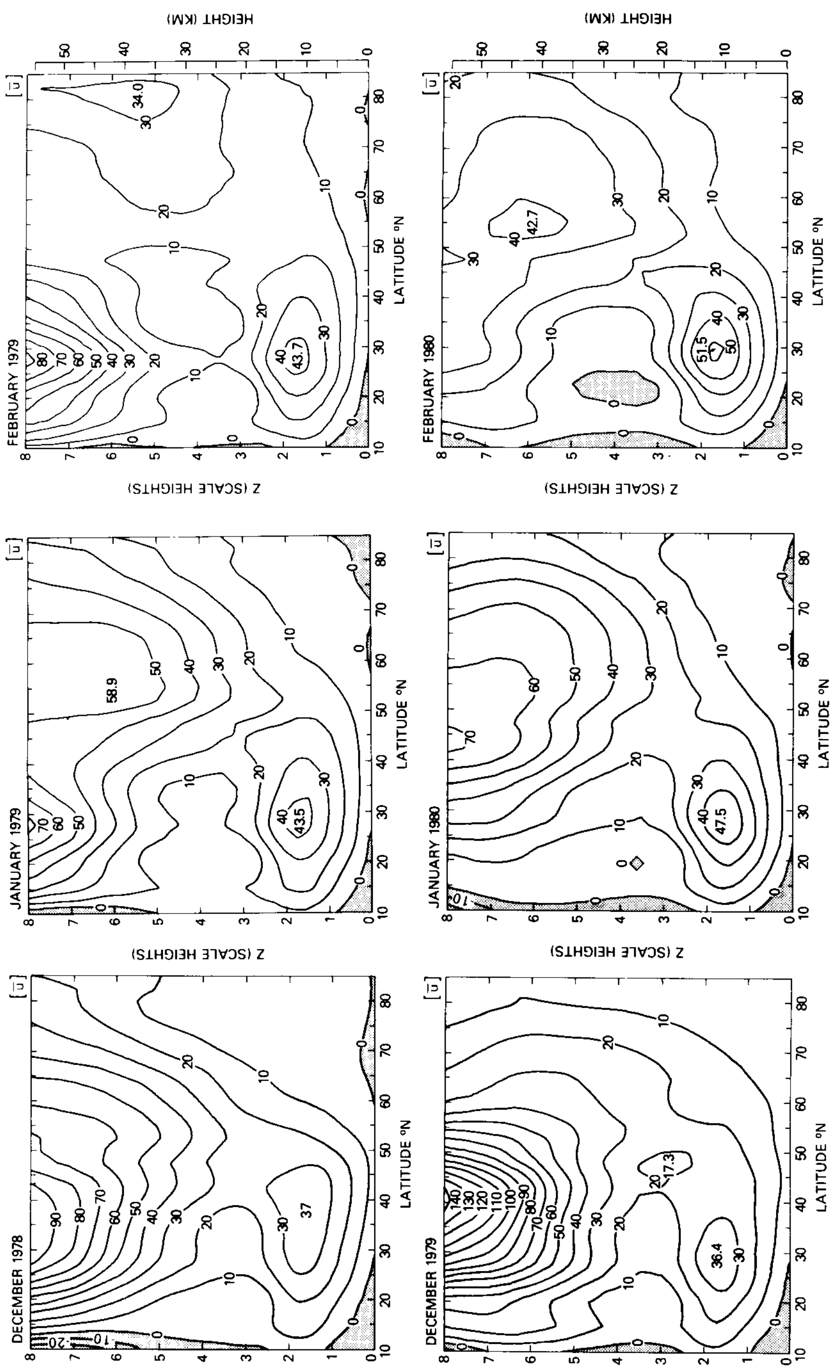
12.5 Interannual Variability in the Stratosphere

It has been known for many years that the long-period variation in the circulation of the equatorial lower stratosphere is dominated not by the annual cycle but by the quasi-biennial oscillation (see Chapter 8). In recent years, with the advent of satellite temperature sounding of the stratosphere it has become clear that the extratropical winter stratosphere also contains considerable variation on interannual time scales. This interannual variability is clearly apparent in the monthly averaged Northern-Hemisphere mean zonal wind sections shown in Fig. 12.5. The figure shows December, January, and February conditions for four separate years. In each year the zonal mean circulation for December is stronger than that for January or February, but there is substantial year-to-year variation in the circulation for each month and in the month-to-month changes.

At present it is not known whether the source of this interannual variability is in the troposphere, the stratosphere, or both. Some idealized models (e.g., Holton and Mass, 1976) have indicated that even in the presence of constant tropospheric forcing, the circulation in the winter stratosphere might oscillate irregularly between two quasi-equilibrium states. In reality, however, the general circulation of the troposphere itself has considerable interannual variability, and the observed interannual variability in the extratropical stratosphere may primarily reflect variability in the tropospheric forcing rather than variability generated in the stratosphere.

12.5.1 The Extratropical QBO

Although the interannual variability in the extratropical stratosphere is much less regular than is the QBO in the equatorial lower stratosphere, a


DECEMBER 1978
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Z (SCALE HEIGHTS)
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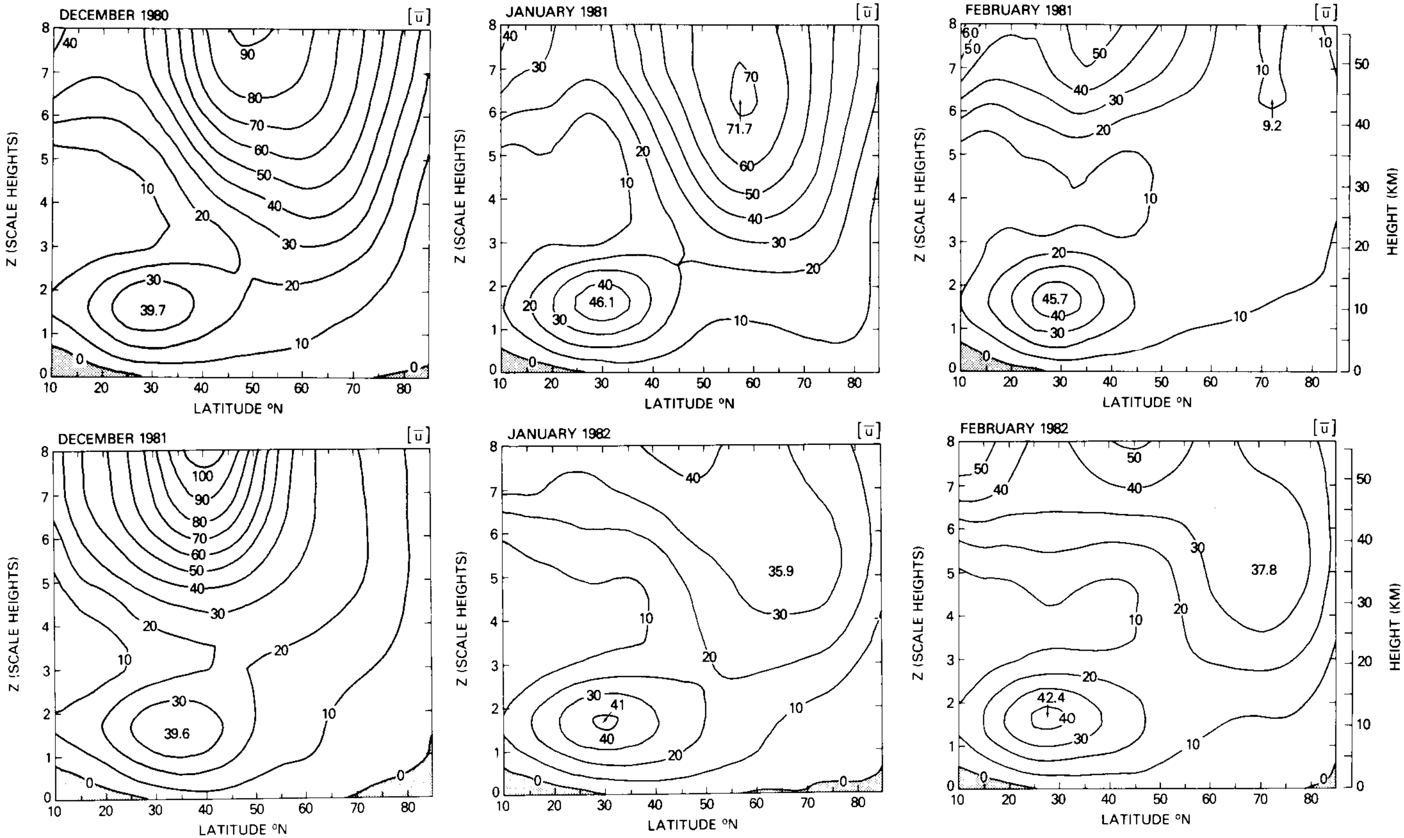

Fig. 12.5. Northern-Hemisphere average December, January, and February mean zonal winds (m s^{-1}) for the four winters of 1978–1979 through 1981–1982. [After Geller *et al.* (1983). American Meteorological Society.]

portion of the extratropical variability does appear to be correlated with the equatorial QBO. Holton and Tan (1980, 1982) composited a 16-year set of gridded monthly mean geopotential height and temperature data for several levels in the stratosphere with respect to the phase of the equatorial QBO. For each of the Northern-Hemisphere winter months they placed the monthly means into either a westerly or an easterly category, depending on the sign of the equatorial zonal wind at 50 mb (see Fig. 8.1). They found that during the westerly phase of the equatorial QBO at 50 mb, the geopotential heights are lower in the polar region and higher in midlatitudes than during the easterly phase (Fig. 12.6). They also found that the stationary planetary wave number 1 had a larger average amplitude during the early winter in years when the equatorial QBO was in its easterly phase at 50 mb. However, little difference in the EP flux pattern was evident in the data. Thus, although it is tempting to speculate that the dynamical connection between the equatorial QBO and the QBO in the polar vortex strength involves differing planetary-wave propagation characteristics due to the QBO in the mean zonal wind profile at low latitudes, there is not much evidence in the data to support such an hypothesis. Rather, observations

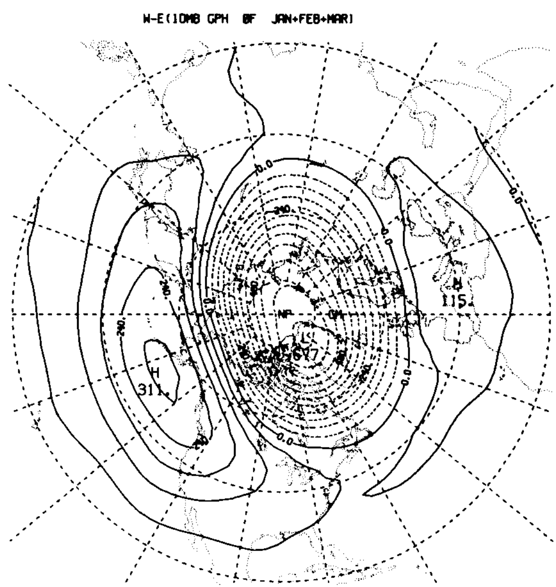

Fig. 12.6. Geopotential height difference (m) at 10 mb for January–March for 16 years of data composited with respect to the phase of the equatorial QBO at 50 mb (westerly category minus easterly category). The outer latitude circle is 20°N. [After Holton and Tan (1982).]

indicate that the extratropical QBO is primarily a zonally symmetric "seesaw" oscillation in which the oscillations in the polar height and thickness fields are out of phase with those in midlatitudes; there is no clear evidence for eddy driving of this seesaw oscillation.

A similar sort of QBO has been observed in the total ozone field. Quasi-biennial oscillations in total ozone have been reported from individual observing stations for many years. Recently the availability of multiyear satellite observations has made it possible to deduce the global pattern of the QBO in total ozone. In the equatorial region, positive ozone deviations occur a few months before the maximum westerlies at 50 mb, consistent with downward advection by the residual circulation associated with the zonal wind QBO (see Fig. 8.5). In midlatitudes the QBO in total ozone appears to have substantial asymmetry in phase between the Northern and Southern Hemispheres, as shown in Fig. 12.7. It should be noted, however,

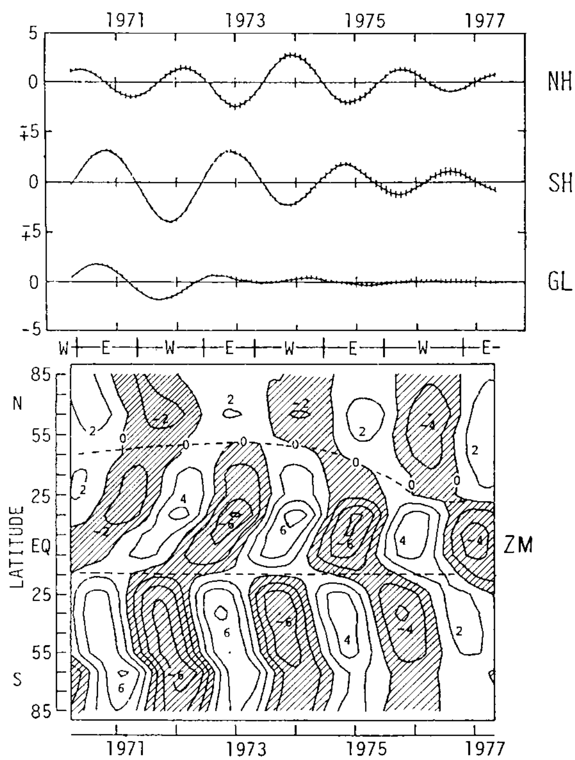

Fig. 12.7. QBO in total ozone (Dobson units) for the Northern Hemisphere (NH), the Southern Hemisphere (SH), global mean (GL), and latitudinally dependent zonal mean (ZM). Easterly (E) and westerly (W) phases of the zonal wind QBO at 50 mb are indicated above ZM. [After Hasebe (1983).]

that much longer data records will be needed to confirm that this phase behavior is an intrinsic aspect of the QBO in total ozone.

12.5.2 *The Influence of Volcanic Emissions*

An important, but highly irregular, source of interannual variability in the stratosphere is the aerosol mass loading that can result from volcanic eruptions. Normally the stratosphere has a thin aerosol layer centered several kilometers above the tropopause, as illustrated in Fig. 12.8. This layer is believed to consist primarily of sulfuric acid particles that result from transport of gaseous carbonyl sulfide (COS) across the tropopause, followed by photochemical conversion to sulfuric acid, nucleation, and particle growth by condensation and coalescence. This background stratospheric aerosol is not believed to play a significant role in the heat budget of the atmosphere.

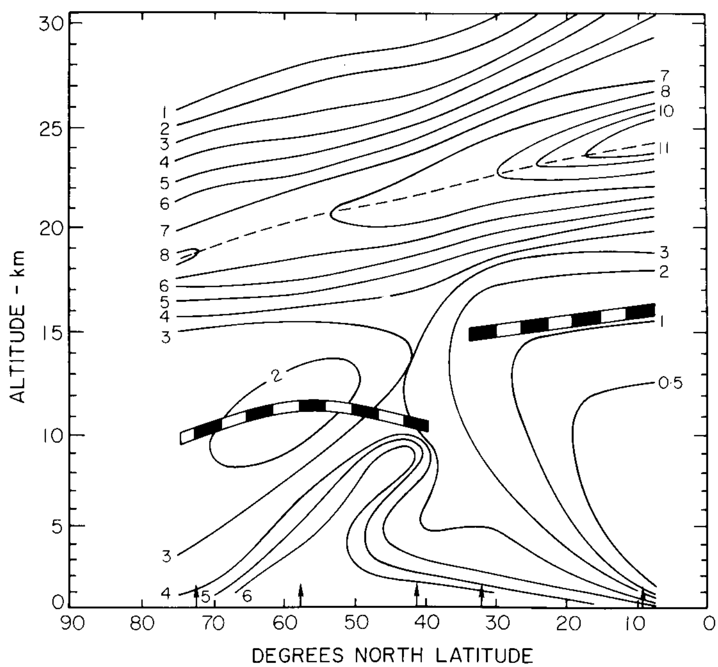

Fig. 12.8. Latitude-height cross section of the background aerosol mixing ratio (particles per milligram of air), June 1973. Heavy broken line marks the conventional tropopause, and dashed line marks the center of the stratospheric aerosol layer. [After Rosen *et al.* (1975). American Meteorological Society.]

Occasionally the stratospheric aerosol density is enhanced by an order of magnitude or more due to massive injections from volcanic eruptions of sulfur gases, which are rapidly converted to sulfuric acid aerosols. The eruptions of Mt. Agung in 1963 and El Chichon in 1982 both caused enormous enhancements in the aerosol loading of the stratosphere. For nearly 6 months following the April eruption of El Chichon the enhanced aerosol was concentrated in a zonal ring at about the 24-km level and in a latitudinal range of about 0° to 30°N. A vertical profile of the lidar scattering ratio (roughly proportional to aerosol mixing ratio) more than 6 months following the eruption is shown in Fig. 12.9. Shibata *et al.* (1984) showed that the very slow vertical spread of the aerosol layer is consistent with a vertical diffusion coefficient of $K_{zz} \approx 10^{-2}\ \mathrm{m^2\ s^{-1}}$, which implies a very low magnitude of turbulent diffusion in the lower-stratospheric easterlies. The lack of significant transport across 30°N is consistent with the weak eddy activity and near radiative equilibrium temperature distribution expected for the extratropical summer stratosphere (see Section 7.2). The absence of cross-equatorial transport may in part be due to the fact that the QBO was in its easterly phase in the lower equatorial stratosphere during this period, so that Kelvin waves should be the dominant equatorial wave mode present. These waves, as discussed in Sections 4.7.1 and 8.3.2, are symmetric about the equator and produce negligible meridional parcel displacements.

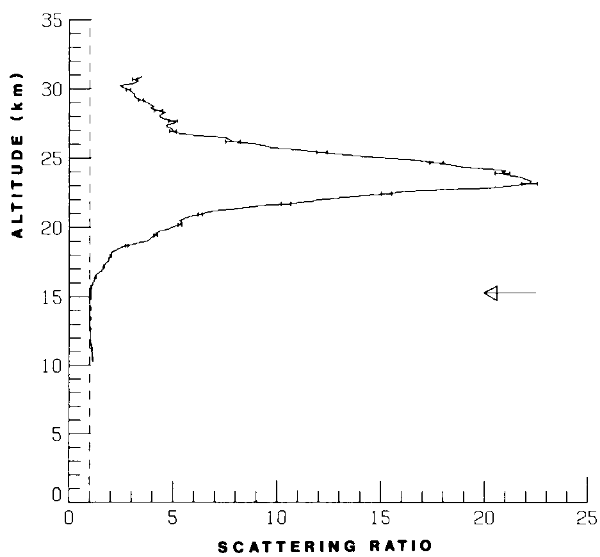

Fig. 12.9. Lidar scattering-ratio profile for October 22, 1982, at 17°N and 82°W. Arrow marks tropopause. [After McCormick and Osborn (1985).]

Such a persistent enhancement in the aerosol layer should influence the radiative equilibrium of the lower stratosphere by scattering solar radiation and by absorption of both solar and terrestrial radiation. For the eruption of El Chichon, radiative-convective equilibrium calculations predict that absorption should raise the radiative equilibrium temperature in the aerosol layer in the lower stratosphere by 3.5 K, while the net effect of the reduced solar radiation and enhanced downward thermal radiation reaching the surface should lead to insignificant temperature changes at the ground, except in the interior of continents (Pollack and Ackerman, 1983). For both the Mt. Agung and El Chichon eruptions, observations (Fig. 12.10) indicate warmings at 30 mb in the tropical belt affected by the aerosol of about 4-6 K. According to Labitzke and Naujokat (1983), these perturbations are more than 3 standard deviations from the 18-year average (1964-1982) at 10°N and 30 mb. Thus, they are clearly associated with the volcanic aerosol layers and do not simply represent normal fluctuations in the strength of the equatorial QBO.

12.5.3 Temperature Trends in the Stratosphere

Because of the coupling between temperature and photochemistry, the existence (or otherwise) of long-term trends in temperature in the stratosphere is a question of some significance. Furthermore, since the summer stratosphere is dynamically undisturbed, and hence has temperatures close to radiative equilibrium, any temperature trend due to the secular increase in CO_2 might be most easily detected in this region. Unfortunately, the rather short period for which reliable stratospheric temperature data are

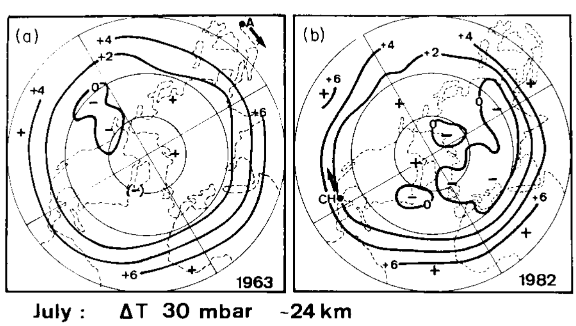

Fig. 12.10. Deviations (K) of the 30-mb temperatures for (a) July 1963 and (b) July 1982 from the 10-year average (1964-1973). The positions of Mt. Agung and El Chichon are shown, together with the mean wind directions for the summers of 1963 and 1982, respectively. [After Labitzke and Naujokat (1983).]

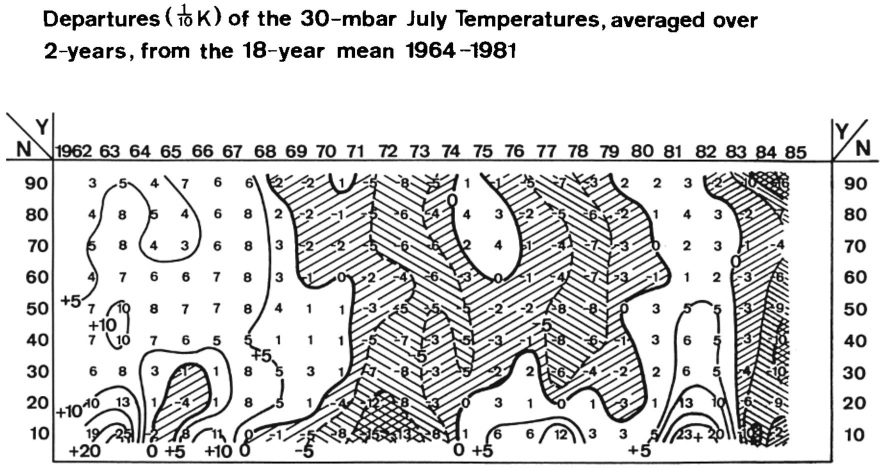

Fig. 12.11. Time-latitude plot of the departures (0.1 K) of the 30-mb July temperatures (averaged over 2 years to remove the QBO) from the 18-year mean of 1964–1981. Negative departures are shaded. [Updated from Labitzke and Naujokat (1983), courtesy of K. Labitzke.]

available, combined with the "red noise" character of natural interannual variability, makes the detection of any such secular trends difficult. The 23-year time series of July mean temperatures at the 30 mb level shown in Fig. 12.11 indicates substantial variation on a decadal timescale. It is not clear, however, whether the overall cooling observed during this period indicates a systematic decrease (as should eventually occur due to the secular increase of CO_2) or merely represents part of a longer-period oscillation. Because of the limited coverage provided by the meteorological radiosonde network, well-calibrated long-term monitoring by satellite is required for definitive detection of global trends in the temperature of the middle atmosphere.

References

12.1. The influence of perturbations in the middle atmosphere on climate is reviewed in Chapter 7 of Brasseur and Solomon (1984).

12.2. Radiatively induced climate perturbations associated with trace gas perturbations are discussed by Ramanathan *et al.* (1985) and by Wang *et al.* (1986). Both papers have extensive bibliographies.

12.3. CO_2 climate change computed in a spectral GCM is discussed by Washington and Meehl (1983).

12.4. Possible stratosphere-troposphere links due to modulations of planetary waves by solar-induced changes in the circulation of the middle atmosphere are analyzed by Schoeberl and Strobel (1978) and Geller and Alpert (1980).

12.5. Interannual variability in the winter stratosphere is discussed by Smith (1983), Geller *et al.* (1984), and Hamilton (1982b). Labitzke (1982) presents an extensive climatology of the interannual variability of the Northern-Hemisphere winter circulation at the 30-mb level and its possible relation to the equatorial QBO. The role of volcanic emissions in stratospheric variability is discussed by Labitzke and Naujokat (1983, 1984).