The Middle Atmosphere: Processes & Interactions

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An overview of the main physical processes and their interactions in the middle atmosphere is presented with emphasis on the interactions. The main physical processes which determine the radiation balance, the thermal structure, the dynamical motions, and the concentrations of ozone and other minor constituents in the stratosphere and mesosphere are first outlined. Then the feedback mechanisms and the multiple interactions amongst the radiative, chemical-photochemical and dynamical processes are described illustratively, but not comprehensively, with special emphasis on the central role of ozone in the stratospheric and lower mesospheric processes.

1 Introduction

There is no rigorous definition of the 'Middle Atmosphere'. Following the widely accepted practice, the Earth's atmosphere between the tropopause and 100 km altitude is considered as the middle atmosphere for the purpose of this overview article. Through conscious planning and coordinated scientific effort on a global scale, the knowledge of the structure, the variability and the physical processes in the middle atmosphere progressed enormously in the last decade and half. This knowledge has enabled the scientists to acquire a clear view of the complexity of interactions between the radiative, dynamical, chemical and photochemical processes in the middle atmosphere; but the understanding of the above interactions is only at the beginning stage, and continuing research is needed in the coming years.

In this overview article, emphasis is placed on the 'interactions' aspect of middle atmosphere processes, because it is this aspect which provides a larger perspective to all scientific studies comprising a large variety of measurements, data analyses and theoretical investigations on specific physical processes.

2 Main Physical Processes

The physical processes in the middle atmosphere can be discussed conveniently in terms of the following four categories:

(i) Radiation—Interactions of solar and terrestrial radiations with the atmospheric constituents.
(ii) Chemistry—Chemical and photochemical reactions and their consequences in the atmosphere.
(iii) Dynamics—Generation, maintenance and dissipation of dynamical motions of various scales, and their effects on other atmospheric processes.
(iv) Structure—Thermal structure and chemical composition of the atmosphere which are in a state of a dynamic equilibrium (or disequilibrium) as a result of processes listed in (i), (ii) and (iii) above.

An outline of the main processes under each category is given below.

2.1 Radiation

Let us begin with an overview of the radiation budget of the Earth-atmosphere system as depicted schematically in Fig. 1(a). Out of the total solar electromagnetic radiation incident at the outer reaches of the Earth's atmosphere, only a minute fraction is absorbed by the atmosphere above 100 km. Without discussing this fraction, let us consider the total solar (e.m.) radiation incident downward at 100 km level as 100 units. Out of this, 3% is absorbed by ozone (mainly in the stratosphere) mostly in the 200-340 nm band of UV radiation. Water vapour, dust and other aerosol particles in the troposphere absorb about 17%, while clouds absorb about 5%, mainly in the visible and infrared bands. Thus about 25% of the incident solar energy is absorbed by various constituents. Backscattering (or reflection) by clouds sends back into space 19% of the incident solar energy, while the backscattering by air, dust and other aerosol particles sends back another 6%, so that 25% of incident solar energy is 'reflected' back to space by the Earth's atmospheric envelope. Out of the remaining 50% reaching the Earth's surface, about 3% is reflected back into space and 47% is absorbed by the Earth-ocean system.

The solar radiation absorbed by the Earth-oceanic surface layers is in turn transferred back upward into the atmosphere in the form of infrared radiation (18%), sensible heat (5%) and latent heat of water vapour (24%) as shown in Fig. 1(b).
It is to be noted from Fig. 1 that only 3% of solar radiation is observed by ozone in the stratosphere; and also the long wave IR radiation emitted predominantly by CO$_2$ (in 15 and 10 $\mu$m bands) and to a small extent by H$_2$O in the stratosphere constitutes only 3% of incident solar energy. The great importance of these apparently small percentages becomes clear if we remember that the atmospheric mass in the stratosphere is slightly less than 10% of the total atmospheric mass. The solar UV radiation in the 200-340 nm band is absorbed by the ozone (O$_3$) in the stratosphere and the resultant heating of the atmospheric air is considerable as shown in Fig. 2. The heating rate maximizes near stratopause, because of the decreasing air density with altitude, though the absolute concentration of O$_3$ (i.e. number of O$_3$ molecules per cm$^3$) maximizes around 27 km height, as shown in Fig. 3. The above heating is balanced by the cooling [Fig. 4(a)] due to the infrared emission (in 15 and 10 $\mu$m bands) by CO$_2$ which is also a trace gas. The resulting net equilibrium temperature in the 30-80 km range constitutes a good approximation for the average temperature structure observed [Fig. 4(b)]. Dynamical effects, however, contribute to superposed temperature fluctuations of different time and spatial scales.

The above radiative equilibrium temperature in the stratosphere and lower mesosphere has a strong latitudinal variation, with highest temperatures near the summer pole and lowest temperatures near the winter pole [Fig. 5(a)]. This pole-to-pole gradient in temperature results in a corresponding gradient in atmospheric pressure and a meridional flow from summer pole towards the winter pole is generated by the pressure gradient force. The Coriolis torque on the meridional flow leads to a geostrophically balanced flow in the zonal direction which is much larger in magnitude than the net meridional flow [Fig. 5(b)]. Following the annual cycle of summer-
winter-summer conditions in each hemisphere, an annual cycle of zonal wind oscillation is generated in the middle atmosphere. This annual cycle is the dominant component of wind motion in the stratosphere and lower mesosphere at latitudes beyond about ±20°. Thus the direct solar radiational heating resulting from the absorption of UV radiation by ozone is the driving source for the major part of dynamical motions in the middle atmosphere at middle latitudes. The central importance of ozone derives partly from this consideration.

Finally, it is to be noted that the interaction of solar radiation with other minor constituents in the middle atmosphere does not cause directly any significant changes in the radiational, thermal or dynamical energy budget of the middle atmosphere. The photochemical reactions of several minor constituents, resulting from their interactions with solar (UV) radiation, play a crucial role in influencing the chemical reactions of ozone and hence its concentration. The resulting changes of ozone in turn cause changes in the temperature, radiation budget and the dynamics of the middle atmosphere. Thus, ozone again plays a central role because of its susceptibility to a large variety of chemical and photochemical reactions with other minor constituents.

The enhanced aerosol concentrations in the stratosphere following volcanic eruptions are now known to cause significant changes in the radiation budget of the stratosphere which lead to significant changes in the stratospheric temperatures. The er-
uption of the El Chichon volcano in Mexico in 1982 deposited an estimated $12 \times 10^6$ tonnes of aerosols into the stratosphere. Satellite-based and ground-based measurements have shown a decrease of ground level temperature due to increased absorption of incoming IR radiation and also an increase of stratospheric temperatures. This is an area in which good measurements and quantitative studies are beginning to emerge. The modelling of aerosol effects in general on the radiation passing through the atmosphere reveals a complex dependence of the absorption and scattering properties of the aerosols on their complex refractive index, size distribution, concentration etc. Quantitative understanding of the aerosol interactions with radiations and their net effect on radiations in the atmosphere requires continuing global scale measurements on aerosols and theoretical studies on their interactions with radiations.

### 2.2 Chemistry

'Chemistry' is used here as a generic term to include photochemical reactions as well. Most of the chemistry of the middle atmosphere is centred around two practical aspects: (i) production and loss of ozone; and (ii) production and loss of ionization in the 60-100 km region.

The chemistry related to ozone ($O_3$) started with the classical scheme suggested by Chapman in 1930:

$$O_3 + h\nu \rightarrow O + O_2$$  
$$O + O_2 + M \rightarrow O_3 + M$$  
$$O_3 + h\nu \rightarrow O + O_2$$  
$$O_3 + O \rightarrow O_2 + O_2$$  
Collisional reaction

The above four reactions provide a simple picture of the production, destruction and the heating role of ozone. While the above scheme continues to be the basic framework for understanding the role of ozone, the major effects of three families of trace species on ozone chemistry (particularly its loss) have been recognized in stages in the last four decades. Bates and Nicolet in 1950 pointed out the importance of $HO_2$ in ozone chemistry; Crutzen and Johnston in 1971 pointed out the important role of the chemistry of $NO_x$ species; Stolarski and Cicerone and Wofsy and McElroy in 1974 pointed out the basic role of $ClO_x$ species in the destruction of ozone. It is now known that there are more than 30 chemical reactions involving more than 35 chemical species which are relevant for determining the production-loss balance of ozone. Without going into the details of the complex chains of chemical reactions, it is possible to get a good feel of the chemistry of ozone destruction from a few examples.

The destruction or removal of ozone is the direct consequence of the chemical reactions of the 'active radicals' with ozone and among themselves. The most important active radicals are:

- $NO$, $NO_2$, $Cl$, $ClO$; and $HO$, $HO_2$

Apart from being highly reactive, the active radicals are usually involved in catalytic reactions, as the following examples would show:

- $NO + O_3 \rightarrow NO_2 + O_2$
- $NO_2 + O \rightarrow NO + O_2$

**Net:** $O + O_3 \rightarrow 2O_2$  
(Removal of odd oxygen)

It is to be noted that $NO$ and $NO_2$ are regenerated and participate repeatedly in the removal of $O_3$ and $O$.

- $Cl + O_3 \rightarrow ClO + O_2$
- $ClO + O \rightarrow Cl + O_2$

**Net:** $O + O_3 \rightarrow 2O_2$

In this reaction cycle also, $Cl$ is regenerated and it is available for the next cycle of reaction. In addition to the above, there are other catalytic cycles, involving the same $NO$, $NO_2$ and $Cl$, $ClO$ pairs, and they are not shown here. Different cycles are effective at different heights due to the dependence of (a) the reaction rates on the ambient pressure and temperature, and (b) the reactions on the concentrations of the species involved. We illustrate this point with the hydrogen-oxygen free radicals $HO$ and $HO_2$:

**Cycle 1:**

- $O + HO \rightarrow O_2 + H$
- $H + O_2 + M \rightarrow HO_2 + M$
- $O + HO_2 \rightarrow HO + O_2$

**Net:** $O + O \rightarrow O_2$

**Cycle 2:**

- $O + HO \rightarrow O_2 + H$
- $H + O_3 \rightarrow HO + O_2$

**Net:** $O + O_3 \rightarrow 2O_2$

**Cycle 3:**

- $HO + O_3 \rightarrow HO_2 + O_2$
- $HO_2 + O \rightarrow HO + O_2$

**Net:** $O + O_3 \rightarrow 2O_2$
Cycle 4: \[ \text{HO} + \text{O}_3 \rightarrow \text{HO}_2 + \text{O}_2 \]
\[ \text{HO}_2 + \text{O}_3 \rightarrow \text{HO} + 2\text{O}_2 \]

Net: \[ \text{O}_3 + \text{O}_3 \rightarrow 3\text{O}_2 \]

Cycles 1 and 2 are of great importance above 40 km in removing \( \text{O} \) and \( \text{O}_3 \); but between 40 and 30 km, cycle 3 dominates as the \( \text{O} \) to \( \text{O}_3 \) ratio decreases; and at 30 km and below cycle 4 becomes most important because it does not utilize \( \text{O} \) (which has a very low concentration) but only \( \text{O}_3 \) (which has a high concentration) at these levels. Similar situation prevails with regard to the other catalytic cycles (not shown here) involving \( \text{Cl} \), \( \text{ClO} \) and \( \text{NO}, \text{NO}_2 \) radicals. The above (and other) 'active radicals' have comparatively short life times because of their highly reactive nature. However, they are generated from more stable 'source' gases. For example, for \( \text{NO} \) and \( \text{NO}_2 \), the source gas in \( \text{N}_2\text{O}_5 \); and for \( \text{Cl}, \text{ClO} \) the source gases are many: \( \text{CFCl}_3, \text{CF}_2\text{Cl}_2, \text{CCl}_4, \text{CH}_2\text{Cl}, \text{CH}_2\text{CCI}_3 \) and other halocarbon compounds. The 'active radicals' finally end up in comparatively stable 'sink' gases which are slowly removed from various altitudes by transport and by photolytic and chemical processes. For example, \( \text{NO}, \text{NO}_2 \) end up in \( \text{HNO}_3, \text{HNO}_4 \), and \( \text{Cl}, \text{ClO} \) end up in \( \text{HCl}, \text{HOCl}, \text{ClNO}_3 \). One can summarize the situation in the following way:

<table>
<thead>
<tr>
<th>Source Gases</th>
<th>Active Radicals (Short-lived)</th>
<th>Sink Gases (Nearly Stable)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(i) ( \text{N}_2\text{O}_5 )</td>
<td>( \text{NO}, \text{NO}_2 )</td>
<td>( \text{HNO}_3, \text{HNO}_4, \text{ClNO}_3 )</td>
</tr>
<tr>
<td>(ii) ( \text{CFCl}_3, \text{CF}_2\text{Cl}_2, \text{CH}_2\text{Cl}, \text{CH}_2\text{CCI}_3 )</td>
<td>( \text{Cl}, \text{ClO} )</td>
<td>( \text{HCl}, \text{HOCl}, \text{ClNO}_3 )</td>
</tr>
<tr>
<td>(iii) ( \text{H}_2\text{O} )</td>
<td>( \text{HO}, \text{HO}_2 )</td>
<td>( \text{HNO}_3, \text{HNO}_4, \text{H}_2\text{O} )</td>
</tr>
</tbody>
</table>

Interestingly, the active radicals are regenerated in some cases from the sink gases either through photolysis or other chemical reactions, as illustrated by the following example:

\( \text{HNO}_3 + h\nu \rightarrow \text{HO} + \text{NO}_2 \)
\( \text{HO} + \text{HCl} \rightarrow \text{H}_2\text{O} + \text{Cl} \)

In the first case, the active radicals \( \text{HO} \) and \( \text{NO}_2 \) are generated from the photolysis of the sink gas \( \text{HNO}_3 \), while in the second case, the radical \( \text{Cl} \) is generated from the sink gas \( \text{HCl} \) through its reaction with \( \text{HO} \). Moreover, the active radicals get interchanged in some interlocked fast reactions as illustrated by the example below:

\( \text{ClO} + \text{NO} \rightarrow \text{Cl} + \text{NO}_2 \)
\( \text{HO}_2 + \text{NO} \rightarrow \text{HO} + \text{NO}_2 \)

In the first case, the active radicals \( \text{ClO} \) and \( \text{NO} \) react chemically and generate the other two radicals \( \text{Cl} \) and \( \text{NO}_2 \), while in the second case a similar process generates \( \text{HO} \) and \( \text{NO}_2 \).

The reactions presented above are only illustrative of the most important of the more than 130 reactions involved directly or indirectly in the ozone chemistry. Considering the fact that a large number (typically 10-20) of simultaneous reactions is taking place at any altitude at any given time\(^4\), the complexity of quantifying the chemistry becomes self-evident. The time rates of the reactions span a wide range: \( 1.8 \times 10^9 \) to \( 2.4 \times 10^7 \) mol. cm\(^{-3}\) s\(^{-1}\) (at 30 km altitude). The reaction rates are very sensitive to the ambient temperature in many cases. The ambient temperature is in turn affected by the absorption of solar radiation by the ozone whose concentration is influenced by the temperature-sensitive reaction rates. Thus, a feedback loop is set up in case of ozone and other trace gases.

All the source gases for the fast reacting radicals are injected into the stratosphere by upward transport from the troposphere. Some of them are generated by natural sources like the atmospheric water vapour \( (\text{H}_2\text{O}) \) and the nitrous oxide \( (\text{N}_2\text{O}) \) which originate in the vegetation and in the biological activities on land and ocean surfaces. Others like the Freons F-11 (\( \text{CFCl}_3 \)) and F-12 (\( \text{CF}_2\text{Cl}_2 \)) are entirely man-made. Continuing release of F-11 and F-12 into the atmosphere (at the rate of about 0.8 millions tons in 1974) was estimated to result in significant reduction of ozone at stratospheric levels, and therefore the use of the above Freons for non-essential purposes like deodorants and hair sprays was banned in some countries and controlled in some other countries. These chlorofluorocarbons (CFCs) are inert and non-reactive. When they reach stratospheric heights, the increased levels of solar UV radiation (\(< 400 \text{ nm}) breaks them up by photolysis into smaller molecules and atoms like \( \text{ClO} \) and \( \text{Cl} \) which are highly reactive. By catalytic reactions like those shown earlier, each of such 'active radical' molecules or atoms can destroy thousands and hundreds of thousands of ozone molecules. Therefore, the presence of the source gases even in extremely small quantities (like a few molecules per trillion molecules of air) can have a significant effect on ozone removal. A reduction of ozone concentrations in the stratosphere can lead to increased levels of UV-B radiation (280-320 nm) reaching the surface of the Earth, with harmful consequences to vegetation, animal life and human health. Because of this practical concern, a vigorous research programme was mounted 15 years ago for the measurements of the various trace gases in the stratosphere and for mo-
delling the chemical-photochemical reactions affecting the ozone concentrations at various levels. References to the literature on the ozone problem can be found in the relevant (review) papers\textsuperscript{15-18}.

Apart from the chemical reactions, transport of trace gases by atmospheric winds in the vertical, meridional and zonal directions leads to their redistributions and concentration changes. For example, ozone production is maximum in the tropical latitudes but the columnar content of ozone maximizes at high latitudes due to meridional transport and smaller loss levels at high latitudes. Thus, wind systems play a major role in influencing the vertical and latitudinal structures of ozone and other minor constituents. Therefore, the dynamical and chemical processes have to be considered together for a quantitative modelling and understanding of the ozone distribution and its role in the atmospheric processes.

2.3 Dynamics

Dynamical motions in the middle atmosphere have a wide range of time scales and spatial scales. A useful classification according to their time scales is as follows.

<table>
<thead>
<tr>
<th>Component of wind</th>
<th>Typical time scale</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Prevailing/Steady wind</td>
<td>Many years</td>
</tr>
<tr>
<td>2. Quasi-biennial oscillation</td>
<td>26 months</td>
</tr>
<tr>
<td>3. Annual oscillation</td>
<td>12 months</td>
</tr>
<tr>
<td>4. Semi-annual oscillation</td>
<td>6 months</td>
</tr>
<tr>
<td>5. Planetary scale waves</td>
<td>2-40 days</td>
</tr>
<tr>
<td>6. Tidal winds</td>
<td>12-24 hr</td>
</tr>
<tr>
<td>7. Gravity waves</td>
<td>5-100 min</td>
</tr>
<tr>
<td>8. Turbulence (small scale)</td>
<td>1-100 s</td>
</tr>
</tbody>
</table>

Thus, the time scales have the wide range of \(10^0\) to \(10^8\) s. The spatial scale, \(L\), of any motion is related to its time scale \(T\), as

\[
L = VT
\]

where \(V\) is the velocity of motion and is typically \(10^{-2}\) to \(10^2\) m/s in the middle atmosphere. Correspondingly the spatial scale, \(L\), has a wide range of \(10^{-2}\) to \(4 \times 10^7\) m.

The above eight components of atmospheric motion are highly non-uniform in altitude and latitude with regard to their relative and absolute average magnitudes. A gross characterization of the relative dominance of different components at different altitudes is given in Fig. 6.

Moreover, the quasi-biennial and semi-annual oscillations of the zonal wind are dominant in the lower and upper levels of middle atmosphere, respectively, only within the tropical latitude belt (say within \(\pm 20^\circ\) of the equator). The annual oscillation of zonal wind becomes dominant in the upper stratosphere and mesosphere at latitudes beyond \(\pm 20^\circ\). Planetary scale waves of different types are present at tropical and extra-tropical latitudes; and they play a very important role in the middle atmosphere. Tidal winds are global, but different modes are important at different latitudes. The ubiquitous gravity waves are present at all latitudes, but the dependence of their spectrum on latitude is poorly known. Turbulence is present everywhere with comparatively large variations in altitude, as turbulence is closely related to the instability condition of the atmosphere.

It is important to note that the physical mechanisms giving rise to different components of wind motion or oscillation are different, though solar radiation absorbed by the Earth-atmosphere system is the basic source of energy for driving all atmospheric motions. For example, the annual oscillation of zonal wind in the upper stratosphere and lower mesosphere is radiatively driven by the latitudinally unequal heating of the stratosphere through the solar UV absorption by ozone. In contrast, the quasi-biennial oscillation and the semi-annual oscillation are dynamically driven by the upward propagating planetary scale equatorial waves originating in the troposphere. These waves deposit their momentum and accelerate the background flow. The planetary scale waves are generated by orographic forcing and thermal forcing in the troposphere. The atmospheric tides are generated from the heating of the relevant atmospheric region due to the absorption of solar radiation by ozone in the stratosphere and by water vapour in the troposphere. The shorter period gravity waves have several possible sources which are yet to be identified on a definitive basis.
Turbulence in the atmosphere results from the unstable condition of a given region and this condition is influenced strongly by atmospheric motions of different scales. In particular, the strong gradients in atmospheric flow associated with any scale of motion are often the main cause of atmospheric instability and turbulence.

The above atmospheric motions of different scales often interact and influence one another, sometimes in a direct way and sometimes in an indirect manner. Such interactions are only known qualitatively so far, and atmospheric research has to go a long way in quantifying these interactions. The importance of atmospheric motions arises from their role as carriers of mass, momentum, energy, heat and the trace gases from one part to another part of the atmospheric medium. Through atmospheric motions, part of the heat, mechanical energy or ozone, generated at one altitude or latitude ends up at another. Therefore, it is necessary to quantify the atmospheric motions of all scales from observations of various types.

In the last 15 years, considerable effort has been made to develop models of the global scale general circulation in the middle atmosphere, taking into account the heat fluxes, the radiative heating/cooling, the chemical reactions, the dynamical motions and their interactions. One result of such a 3-D model with regard to the meridional circulation is shown in Fig. 7. In fact, in Fig. 7 the meridional flow lines in the entire atmosphere from surface to 70 km height are shown. There are three distinct circulation cells in the troposphere in each hemisphere; but in the stratosphere there are only three cells on the global scale, one of them being rather limited in extent. At mesospheric levels, there are only meridional cells from pole to pole. It is to be noted further that each cell consists of a predominantly rising motion in a particular narrow latitude range followed by a predominantly horizontal (northward or southward) motion over a large range of latitudes, then a predominantly sinking motion in a narrow latitude range and finally a horizontal meridional flow at a lower height in a direction opposite to that at the higher level. From the consideration of mass continuity along the flow lines, it is clear that the vertical velocities are much smaller (a few mm/s to a few cm/s) than the horizontal velocities (several m/s) because the vertical path lengths of the cells are 2-4 orders of magnitude smaller than the horizontal path lengths. These meridional circulation cells, together with the zonal winds of larger magnitude, keep the entire atmosphere well-mixed up to about 100 km and consequently the concentrations of the stable trace gases like CO₂ have uniform values everywhere (as percentages of the ambient air). Finally, the above circulation cells play a vital role in transporting, vertically and latitudinally, the trace gases from their (sometimes limited) source regions to distant regions of the atmosphere.

2.4 Structure

The structure of the middle atmosphere has three basic aspects: (a) temperature (b) winds and (c) chemical composition. The average pressure and density of the atmosphere decrease exponentially with altitude, if we ignore the perturbations. The mean temperature has an altitude structure (shown in Fig. 4) which has significant changes with latitude, especially at mesospheric levels; and its seasonal changes are significant at higher latitudes particularly in the mesosphere. More important is the fact that shorter time scale mesospheric temperature changes are caused predominantly by dynamical effects which are understood only in qualitative terms so far. This is an aspect which needs a great deal of experimental and theoretical investigations.

The gross characteristics of winds in the middle atmosphere are known. But their variability is quite large and we do not have an adequate observational knowledge of this variability. For example, a recent analysis has shown that even long period oscillations like the annual oscillation have a large longitudinal asymmetry. The observations of the planetary scale waves in the middle atmosphere are still meagre. On the whole, our observational knowledge of the winds in the middle atmosphere is far from satisfactory especially with regard to their variability.

The chemical composition of middle atmosphere has two parts: one part comprises the very stable...
gases with uniform concentration (in terms of their percentages by volume) throughout the atmosphere up to the turbopause around 100 km, and the other part consists of all the minor constituents or trace gases which are highly non-uniform in their distributions with altitude and latitude with large variations in time as well. The most important of both categories of atmospheric gases are listed in Table 1.

3 Interactions

The radiative, photochemical, chemical and dynamical processes in the middle atmosphere interact with one another all the time. The ever changing structure of any part of the middle atmosphere is the consequence of such interactions. A broad view of the nature of these interactions is given with reference to the stratosphere, because our understanding of such interactions in the stratosphere has attained a good degree of clarity qualitatively at least.

Fig. 8 shows schematically the various processes and their interconnectedness. Let us begin with photochemical, and chemical processes. Solar radiation dissociates \( \text{O}_2 \) molecule into \( \text{O} \) atoms which combine, through 3-body collision, with \( \text{O}_2 \) to produce \( \text{O}_3 \). \( \text{O}_3 \) gets destroyed in several reactions including collisional interaction with \( \text{O}_2 \) and fast chemical reactions with active radicals like \( \text{OH}, \ \text{HO}_2, \ \text{NO}, \ \text{NO}_2, \ \text{Cl}, \ \text{ClO} \). The above radicals are generated from chemically stable source gases originating at the Earth’s surface and transported upward through the troposphere. Some of the source gases are natural (e.g. \( \text{N}_2, \text{H}_2 \text{O} \)) and some are man-made (e.g. \( \text{CFC}_1, \text{CF}_2\text{Cl}_2 \)). Photolysis and photochemical reactions of the source gases in the stratosphere, where the UV radiation intensity is sufficient to cause such reactions, generate the active radicals. The reactions of the radicals with \( \text{O}_3 \), with \( \text{O} \) and among themselves are highly localized in the region of their production because their reaction times \( (T_R) \) are extremely small compared to their transport time \( (T_t) \). Their reactions lead to the formation of the ‘reservoir’ species like \( \text{N}_2\text{O}_3 \) and \( \text{H}_2\text{O}_2 \) with life times of several hours which leads to some transport effects to be experienced by such species. The ‘reservoir’ species regenerate the radicals in some reactions and they also get converted to the more stable sink gas species like \( \text{HNO}_3, \text{HCl}, \text{H}_2\text{O}, \text{CINO}_3 \). These stable sink species get transported slowly (through meridional circulation cells) into the troposphere where they get washed down to the Earth’s surface by precipitation.

In the above scheme of chemical and photochemical reactions, ozone plays a central role. The increase of the source trace gases in the stratosphere results in the increase of the radicals which cause the increased loss of ozone. The loss is a complicated function of source gas concentrations injected into the stratosphere and it is certainly height dependent. Estimates of the loss due to the known levels of \( \text{CFC}_1, \text{CF}_2\text{Cl}_1 \) injections vary from a few per cent to a few tens of per cent. Modelling studies on this are continuing. Though there are some reactions among the trace gases without involving ozone, the effects of such reactions on stratospheric structure are negligible, and so are the effects on the short wave or long wave radiation passing through the middle atmosphere. Ozone is the principal agent causing significant changes in the radiation, temperature and dynamical motions in the stratosphere through the changes in its concentration which are in turn generated by the chemical, photochemical reactions involving various trace gases.

The absorption of the solar UV radiation by ozone has two consequences: (i) it causes heating of the
ambient gas and thus effects the thermal structure and (ii) O₃ and O₂ cut off completely the harmful UV radiation at wavelengths below 280 nm, and O₃ reduces it considerably in the 280-340 band, as the radiation penetrates downward to the Earth's surface. In the absence of this cut-off and reduction, plant and animal life on the earth (including the oceans) would suffer immense damage. In fact, most life forms can get extinguished.

The heating of the ambient gas is the driving source for the meridional circulations and zonal winds in the stratosphere. Winds are generated from temperature gradients but they also tend to reduce the temperature gradients as they convect heat from higher temperature to lower temperature regions, and a dynamic equilibrium is usually maintained. Thus the changes in ozone can cause changes in temperature and winds which in turn interact with each other. Ozone concentration itself is affected by winds and temperature: (i) Winds transport ozone leading to its redistribution and changes of its concentration and this process is very effective in the lower stratosphere (where O₃ concentration is maximum) because the reaction times of O₃ chemistry there are larger than the transport time; (ii) The temperature affects O₃ concentration through the temperature dependence of the rates of chemical reactions which destroy ozone; it turns out that the ozone concentration is inversely proportional to temperature in the upper stratosphere and lower mesosphere where the chemical radiative equilibrium prevails due to the reaction times of O₃ reactions, being small compared to the transport time. Consequently the temperature changes in the lower stratosphere and upper stratosphere will have a different phase relationship with respect to the ozone changes.

The above description brings out the interactive nature of the radiative, chemical-photochemical, thermal and dynamical processes centred around ozone in the stratosphere. The above interactive processes are also affected to a lesser extent by other minor constituents like aerosols, water vapour and carbon dioxide as shown in Fig. 8. Aerosols, H₂O and CO₂ absorb the incoming solar radiation and the outgoing long wave (IR) radiation while H₂O and CO₂ also emit infrared radiation; and aerosols and O₂ also scatter radiation. The emission of 15 μm (and 10 μm) radiation by CO₂ is the principal mechanism of heat loss in the stratosphere. A much smaller fraction of heat loss is caused by the 9.6 μm emission by O₃. The cooling rates due to the above emissions by CO₂ and O₃ together are shown already for various altitudes in Fig. 4(a). The percentage concentration of CO₂ is uniformly distributed in the atmosphere but it is now known to be undergoing a steady increase with time. The estimated consequence is a global warming up at the
Earth's surface level and cooling in the stratosphere\textsuperscript{21}. If this trend continues, it may have serious consequences for the global climate. The recent discovery of ozone 'hole' (i.e. depletion of ozone concentration) over Antarctica and the lack of a satisfactory explanation for it illustrate dramatically the meagerness of our knowledge of middle atmosphere processes and the need for a concentrated research effort in the coming years\textsuperscript{22}.

Finally, the middle atmosphere is coupled to the troposphere very closely\textsuperscript{23}. The trace gases which play such a vital role in the stratosphere are transported upward from the troposphere through meridional circulation cells and dynamical changes in the height of the tropopause which often acts as a cold trap for the trace gases because of the \(-70^\circ\text{C}\) to \(-80^\circ\text{C}\) level temperatures there. Mass exchange also takes place between the troposphere and stratosphere. The waves generated in the troposphere propagate upward into the stratosphere and mesosphere, and deposit substantial part of their momentum and energy there through wave mean-flow interactions and dissipative processes. The Earth-emitted infrared radiation is absorbed and scattered by \(\text{CO}_2\) and \(\text{H}_2\text{O}\) in the stratosphere; and the levels of this infrared radiation reaching the stratosphere are modulated by the cloud cover and the water vapour content in the troposphere. The UV and IR solar radiation reaching the troposphere is in turn affected by the concentration levels of \(\text{O}_3\), \(\text{CO}_2\), \(\text{H}_2\text{O}\) and aerosols in the stratosphere.

The interactions and coupling processes outlined above are well recognized now. But a quantitative understanding of the coupling processes can only be gained through sustained long term effort on a global scale in measurements, data analyses, theoretical studies, and modelling.

In this overview presentation, some important aspects like the ionization and D-region chemistry\textsuperscript{24}, the role of aerosols, the electrical processes in the middle atmosphere and the stratospheric warmings have not been presented. The choice is deliberate.

The attempt has been to communicate a clear view of the complex interactions and feedback mechanisms amongst the various physical processes through an illustrative (rather than an exhaustive) presentation of the main elements.

References