Chapter 1
Structure and Composition of the Lower and Middle Atmosphere

1.1 The Evolution of the Earth's Atmosphere

The history of the Earth's atmosphere prior to one billion years ago is not clearly known. Scientists have studied fossils and made chemical analysis of rocks to find out how life on Earth evolved to its present form. Several theories have been suggested. It is hypothesized that life developed in two phases over billions of years.

In the first phase, explosions of dying stars scattered through the galaxy and created swirling clouds of dust particles and hot gases. These extended trillions of kilometers across space. As the cloud cooled, fragmented small particles adhered to each other. Over 4 billion years ago, the cloud had formed into a flattened and slowly rotating disk. The Sun was born in the center of this disk. Away from the disk, Earth and the other planets formed as tiny pieces of matter which were drawn together. Earth started out as a molten mass that did not cool for millions of years. As it cooled it formed a thin, hard crust with no atmosphere or oceans.

Molten rock frequently exploded through the crust. Water vapor was released from the breakdown of rocks during volcanic eruptions. Eventually, the crust cooled enough for this vapor to condense and come down as rain to form the oceans that covered most part of the Earth.

In the second phase, scientists have hypothesized that bubbles floating on the ancient ocean trapped carbon-containing molecules and other chemicals essential for life. These bubbles may have burst and released these chemicals into the atmosphere. Organic compounds formed and dissolved in the early atmosphere, collecting in the shallow waters of the Earth. However, no one is aware how the first living cells developed between 3.6 and 3.8 billion years ago. Eventually, these protocells developed into cells having the properties presently known as life.

These unicellular bacteria multiplied in the warm shallow waters, where they mutated and developed into a variety of protests and fungi. About 600 million years ago plants and animals were formed on the Earth. Life could not develop then on land since there was no ozone layer to shield early life from damaging ultraviolet (UV) radiation. The photosynthetic bacteria which emerged about 2.3–2.5 billion

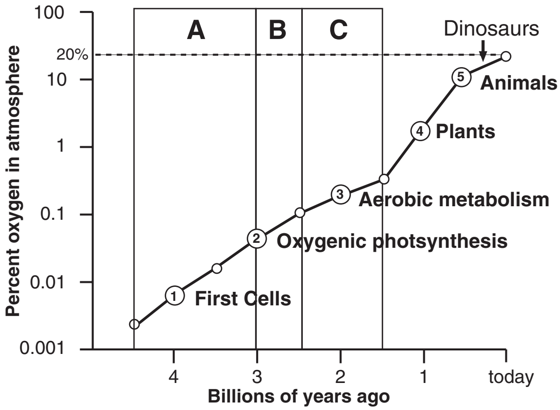

Fig. 1.1 Evolution of the concentration of oxygen in the Earth's atmosphere: (**A**) No oxygen produced by the biosphere; (**B**) Oxygen produced but absorbed by Oceans and sea bed rock; and (**C**) Oxygen absorbed by land surfaces and formation of ozone layer (Courtesy: Tameera, Wikipedia)

years ago, could remove carbon dioxide (CO_2) from the atmosphere and, using sunlight, combine it with water to make carbohydrates. In the process they created oxygen (O_2) and released it into the ocean. Some of the oxygen escaped into the atmosphere.

The evolution of geological and biological events leading to changes of oxygen contents in the Earth's atmosphere since the formation of the planet are shown in Fig. 1.1. Evidence of oxygen in the ocean is noted between 2.5 and 3 billion years ago, by identifying oxidized iron bands in the seabed rock. Ozone layer formation was substantiated from the oxidized iron bands in the land 1.5–2.5 billion years ago. Biological events show that the photosynthetic bacteria started producing oxygen 3 billion years ago and the aerobic metabolism evolved about 2 billion years ago. Evolution of multicellular plants and animals started in the later parts as illustrated in Fig. 1.1.

Earth's atmosphere was formed over a period of 2 billion years. Some of the oxygen was converted into ozone (O_3), which was produced in the lower stratosphere and protected life from harmful UV radiation. This allowed green plants to live closer to the surface of the ocean, making it easier for oxygen to escape into the atmosphere. About 400–500 million years ago the first plants began to exist on land. Over the following millions of years a variety of land plants and animals evolved. As more plants appeared, the levels of oxygen increased significantly, whereas the carbon dioxide levels dropped. At first it combined with various elements, such as iron, but eventually oxygen accumulated in the atmosphere resulting in mass extinctions and further evolution. With the appearance of an ozone layer life-forms were better protected from UV radiation. The present nitrogen-oxygen enriched atmosphere is sometimes referred to as Earth's third atmosphere, in order to distinguish the current chemical composition from two notably different previous compositions.

1.1.1 Living Earth

Earth is really a wonderful planet. It is the only planet in our solar system that has all the necessary components to support life. Our Earth is only a tiny part of the universe, but it is the home of human beings and many other organisms. These organisms can live on Earth because it has an atmosphere. Animals and plants live almost everywhere on the surface of Earth.

The atmosphere moderates daytime and nighttime temperature swings. The atmosphere filters radiant energy during the day, preventing the surface from overheating. At night the atmosphere prevents a large part of the radiant heat from escaping back into space, keeping the surface warm. Most organisms, both plants and animals, essentially need water to live. Seventy-one percent of the Earth's surface is covered by water. Living things also need nitrogen, oxygen, and carbon dioxide. Earth's thin layer of atmosphere provides all of these elements. The atmosphere also screens out lethal levels of the Sun's UV radiation. The atmosphere, however, could not exist if Earth were not at the existing distance from the Sun.

1.2 Earth's Atmosphere and Its Composition

The gaseous envelope surrounding the Earth is known as its atmosphere. It is a relatively stable mixture of several types of gases from different origins. It has a mass of about 5.15×10^{15} tons held to the planet by gravitational attraction. The mean molecular mass of air is 28.966 g mol^{-1}.

The atmosphere is a mixture of gases, some with fairly constant concentrations, others that are variable in space and time. In addition, there are suspended particles (e.g. aerosol, smoke, ash, etc.) and hydrometeors (e.g. cloud droplets, raindrops, snow, ice crystals, etc.). Table 1.1 shows the composition of dry air in the Earth's atmosphere below 100 km.

Nitrogen, oxygen, and argon account for about 99.96% of the permanent gases. Of the variable constituents, carbon dioxide can be somewhat variable in concentration on a localized basis at low levels. Water vapor is a highly variable constituent, with concentrations ranging from nearly zero in the coldest and dry regions of the atmosphere up to as much as 4% by volume in hot and humid air masses. Ozone, the other major greenhouse gas, also varies distinctly. In addition to these variable constituents there are also aerosols and hydrometeors which can vary widely in space and time.

The proportions of gases, excluding water vapor and ozone, are nearly uniform up to a height of about 100 km above the Earth's surface. Eddies are effective at mixing gases in the lowest 100 km of the atmosphere. This part of the atmosphere is called *homosphere*. More than 99.9% of the total atmospheric mass, however, is concentrated in the first 50 km from Earth's surface.

Table 1.1 Major constituents of the Earth's atmosphere upto 100 km (dry air)

Constituents	Percentage by volume	Molecular weight (g mol^{-1})
(A) *Constant concentrations*		
Nitrogen (N_2)	78.08	28.01
Oxygen (O_2)	20.95	32.00
Argon (Ar)	0.933	39.95
Carbon dioxide (CO_2)	0.033	44.01
Neon (Ne)	18.2×10^{-4}	20.18
Helium (He)	5.2×10^{-4}	04.02
Krypton (Kr)	1.1×10^{-4}	83.80
Xenon (Xe)	0.089×10^{-4}	131.29
Hydrogen (H)	0.5×10^{-4}	02.02
Methane (CH_4)	1.5×10^{-4}	16.04
Nitrous oxide (N_2O)	0.27×10^{-4}	44.01
Carbon monoxide (CO)	0.19×10^{-4}	28.01
(B) *Variable concentrations*		
Water vapor (H_2O)	0–4	18.02
Ozone (O_3)	$0–4 \times 10^{-4}$	48.02
Ammonia (NH_4)	0.004×10^{-4}	17.02
Sulphur dioxide (SO_2)	0.001×10^{-4}	64.06
Nitrogen dioxide (NO_2)	0.001×10^{-4}	46.05
Other gases	Trace amounts	—
Aerosol, dust, gases	Highly variable	—

Nitrogen and oxygen make up to 99% of the atmosphere at sea level, with the remainder comprising CO_2, noble gases, and traces of many gaseous substances. Commonly, the unit of concentration used when referring to trace substances is parts per million (ppm). The volume fraction is equal to the mole fraction. Hence, a mole fraction of 3.55×10^{-4} for CO_2 is equal to 355 ppm. There is increasing evidence that the percentages of trace gases are changing because of both natural and human factors. Carbon dioxide, nitrous oxide, and methane (CH_4) are produced by the burning of fossil fuels, expelled from living and dead biomass, and released by the metabolic processes of microorganisms in the soil, wetlands, and oceans. Atmospheric temperature and chemistry are generally influenced by the trace gases.

Some of the gases do not have uniform mixing ratios (e.g. ozone, water vapour, etc.) in the lowest 100 km. They can have a source at the surface or in the atmosphere, or a sink at the surface or in the atmosphere. If the gas's lifetime is shorter than the time it takes to get transported from one place to another, then the gas may not be uniformly distributed throughout the atmosphere.

Above 100 km, the mixing of air parcels is dominated by molecular diffusion. This part of the atmosphere is subjected to bombardment by radiation and high-energy particles from the Sun and outer space. This barrage has profound chemical effects on the composition of the atmosphere, especially the outer layers. In addition, gaseous molecules are influenced by gravity, leading to lighter molecules being found in the outer layers and heavier molecules closer to the Earth. Consequently, the composition of the atmosphere beyond the middle atmosphere is not uniform and is known as *heterosphere*. The upper boundary at which gases disperse into space lies at an altitude of approximately 1,000 km above sea level.

1.2.1 Formation of Homosphere and Heterosphere

Formation of the homosphere and heterosphere in the atmosphere can be explained as follows. Based on the concept of diffusion, the vertical molecular flux is given as

$$F = -D(\nabla N) \tag{1.1}$$

where D is the diffusion coefficient and N is the molecular number density. The negative sign indicates that the fluxes by diffusion are down-gradient.

The rate of change of N with respect to time is given by

$$\frac{\partial N}{\partial t} = -D(\nabla N) = D\nabla^2 N \tag{1.2}$$

The vertical molecular flux in terms of the rate of change of N with height is

$$F = -D\frac{\partial N}{\partial z} \tag{1.3}$$

The diffusion caused by eddies can be estimated by:

$$F_{\text{eddy}} = -K_{\text{eddy}}\frac{\partial N}{\partial z} \tag{1.4}$$

The solution to this diffusion equation for atmospheric conditions results in a time scale for diffusive travel, as time equals $< X^2 > /2D$.

Now, let us use kinetic theory to understand the molecular diffusion coefficient D. The resulting expression for D is:

$$D = k_{\text{B}}\frac{(T)^{3/2}}{(p)(1/m)^{1/2}} \tag{1.5}$$

where k_B is the Boltzmann's constant (1.38×10^{-23} J K^{-1}), T is the temperature, p is the pressure, and m is the molecular mass. At the surface, $D = 2 \times 10^5$ m^2 s^{-1}. When $p = 5 \times 10^{-7}$ atm, T is constant, $D \sim 40$ m s^{-1}.

When eddy diffusion dominates over molecular diffusion, the gases are well mixed and form the homosphere. When molecular diffusion dominates over eddy diffusion, the gases separate according to mass, as we see in the heterosphere.

1.3 Atmospheric Pressure

Atmospheric pressure is the pressure above any area of the Earth's atmosphere caused by the the weight of the air. The air pressure varies spatially and temporally, because the amount and weight of air above the Earth vary with location and time. Atmospheric pressure shows a semidiurnal variation caused by global atmospheric tides. The tidal effect is stronger in tropical zones and almost absent in polar regions. The average atmospheric pressure at sea level is about 1,013.25 hectapascals (hPa).

1.3.1 Vertical Structure of Pressure and Density

Meteorological parameters, such as pressure and air density, vary dramatically with height in the atmosphere. The variation can be over many orders of magnitude and is very much larger than horizontal or temporal variations. It is therefore necessary to define a standard atmosphere in which geophysical quantities have been averaged horizontally and in time, and which vary as a function of height only. The quasi-exponential height dependence of pressure and density can be inferred from the fact that the observed vertical profiles of pressure and density on the semilog plots closely resemble straight lines.

The standard atmospheric values specified by the International Civil Aviation Organisation (ICAO) are: (i) sea level pressure (p) is 1,013.2 hPa; (ii) sea level density (ρ) is 1.225 kg m^{-3}; (iii) sea level temperature (T) is 288.15 K; and fixed lapse rates for p and T.

The vertical variation of pressure (p) with height (z) may be derived as approximately (see Wallace and Hobbs 2006) as:

$$p(z) = p(0)\exp\left(\frac{-z}{H}\right) \tag{1.6}$$

where $p(z)$ is the pressure at height z above sea level, $p(0)$ is the sea level pressure, and H is a constant called the scale height of the atmosphere. Pressure decreases by a factor of e in passing upward through a layer of depth H. For the Earth's atmosphere, H is about 8.4 km. This equation is valid only for an isothermal atmosphere, in which the temperature remains constant with height.

A similar approximate expression may be derived for density ρ as follows:

$$\rho(z) = \rho(0)\exp\left(-\frac{z}{H}\right) \tag{1.7}$$

Note that density also decreases rapidly with height. It can be shown that half of the mass of the Earth's atmosphere is below the 500 hPa level or an altitude of about 5.5 km.

Figure 1.2 illustrates the vertical profiles of pressure in the troposphere and stratosphere. As elevation increases, fewer air particles are above. Therefore the

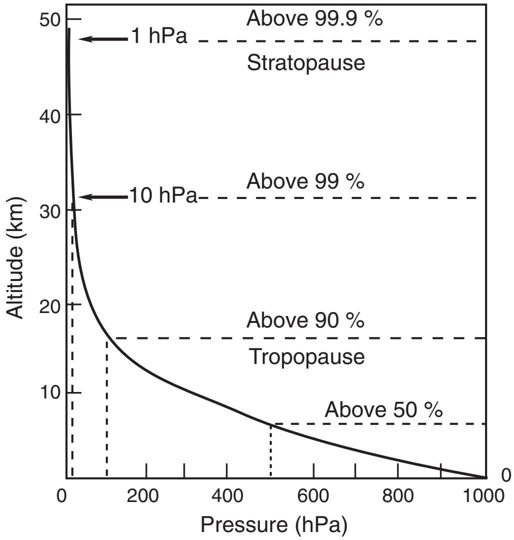

Fig. 1.2 Vertical distribution of atmospheric pressure (US Standard Atmosphere)

atmospheric pressure decreases quasi-exponentially with increasing altitude. The atmospheric pressure drops by $\sim$50% at a height of about 5 km above the Earth's surface (see Fig. 1.2). At an altitude of 50 km the pressure (i.e., mass of particles above unit area at that level) is about 1 hPa so that only about 0.1% of the mass of the atmosphere lies above that level. Similarly, because the pressure at 90 km is about 0.001 hPa, only about one millionth of the mass of the atmosphere lies above that level.

1.4 Atmospheric Thermal Structure

The atmospheric layers of the Earth are characterized by variations in temperature produced by differences in the radiative and chemical composition of the atmosphere at different altitudes. With increasing distance from Earth's surface the chemical composition of air becomes progressively more dependent on altitude and the atmosphere becomes enriched with lighter gases. Based on the temperature changes with height, the Earth's atmosphere is divided into mainly four concentric spherical strata by narrow transition zones. Each layer is a region where the change in temperature with respect to height has a constant trend. The layers are called *spheres* and the transition zones between concentric layers are called *pauses*. The four concentric layers of the atmosphere are termed as troposphere, stratosphere, mesosphere, and thermosphere. The vertical distribution of temperature in the Earth's atmosphere is shown in Fig. 1.3.

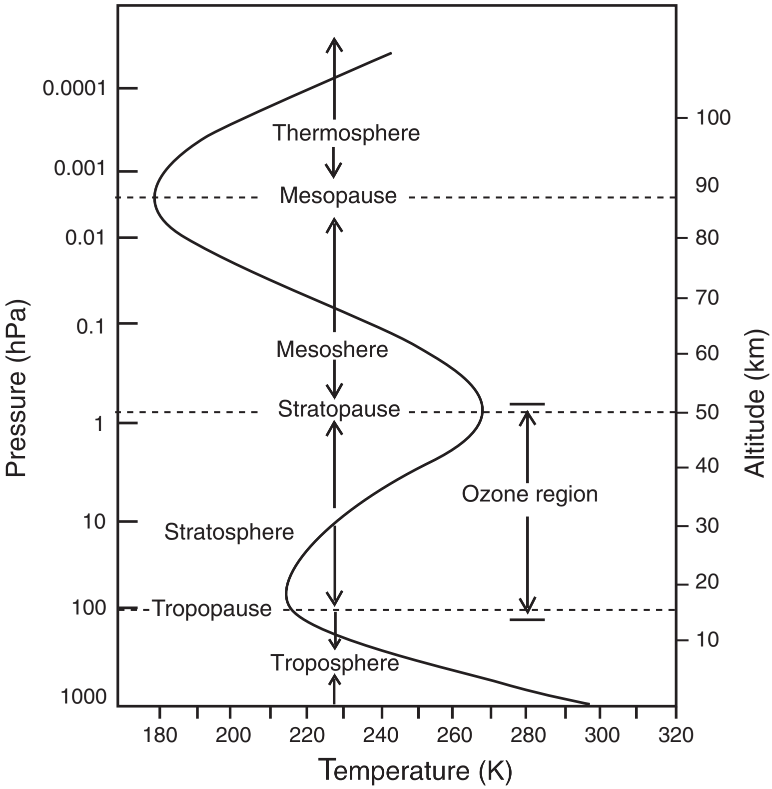

Fig. 1.3 Vertical thermal structure of Earth's atmosphere upto 120 km (Adapted from G. Brasseur and S. Solomon 1984)

1.4.1 Troposphere

Troposphere is the lowest part of the atmosphere and is closer to the Earth, and extends about 8 km above the poles and 18 km over the equator. It is the densest part of the atmosphere which contains almost all the water vapor, clouds, and precipitation. Temperature generally decreases with height in the troposphere at about 6–7°C km^{-1} in the lower half and 7–8°C km^{-1} in the upper half. Because of the general decrease of temperature with height and the presence of weather systems, the troposphere is often characterized by fairly significant localized vertical motions, although these are generally much smaller than horizontal motions. Sometimes shallow layers may be present in the troposphere in which temperature increases with height. These *inversions* inhibit vertical motion. The presence of water vapor, clouds, storms, and weather, contributes to the significance of the troposphere.

Water vapor plays a major role in regulating air temperature because it absorbs solar energy and thermal radiation from the planet's surface. The troposphere contains 99% of the water vapor in the atmosphere. The water vapor content, however, decreases rapidly with altitude, thus reflecting the change in temperature. Water vapor concentrations also vary with latitudinal position, being greatest above the tropics and decreasing toward the polar regions. All weather phenomena occur within the troposphere, although turbulence may extend into the lower portion of the stratosphere. Troposphere means region of *turning* or *mixing*, and is so named because of vigorous convective air currents within the layer.

The troposphere is bounded at the top by the *tropopause*, whose altitude varies considerably depending on the location and type of weather systems, latitude, etc. The temperature and altitude of the tropopause at a given location can vary rapidly depending on prevalent weather systems.

The tropopause may be considered as the base of a large inversion layer, i.e., the stratosphere, which inhibits vertical mixing. Consequently there are often significant concentration gradients across the tropopause. For example, the concentration of water vapor, which results largely from evaporation from the Earth's surface, decreases distinctly above the tropopause while ozone concentration increases noticeably. The moist ozone-poor tropospheric air does not mix much with the dry, ozone-rich stratospheric air.

1.4.2 Stratosphere

The stratosphere is the second major stratum in the atmosphere. It resides above the tropopause upto 50 km as shown in Fig. 1.4. The air temperature in the stratosphere increases gradually to around 273 K at the stratopause ($\sim$50 km), which is marked by a reversal in the temperature trend. Because the air temperature in the stratosphere increases with altitude, it does not cause convection and has a stabilizing effect on atmospheric conditions in the region and confines turbulence to the

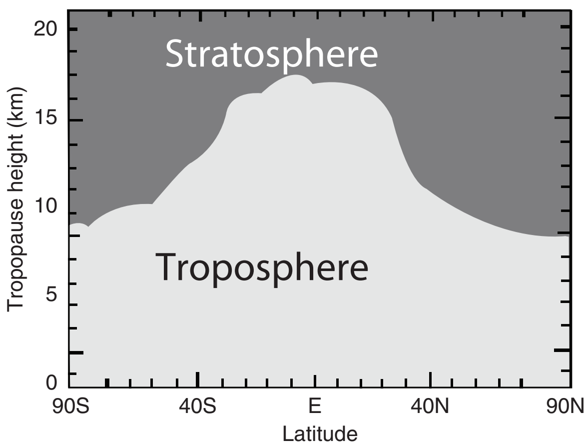

Fig. 1.4 Latitudinal variation of the vertical extension of troposphere and lower stratosphere (Adapted from B. Geerts and E. Linacre)

troposphere. As water vapor content within the stratosphere is very low, ozone plays the major role in regulating the thermal regime of this layer. Temperature increases with ozone concentration. Solar energy is converted to kinetic energy when ozone molecules absorb ultraviolet radiation, resulting in heating of the stratosphere.

The vertical temperature gradient in the stratosphere strongly inhibits vertical mixing, in contrast to the situation in the troposphere. From about 20 to 32 km there is usually a near-isothermal layer, whereas the temperature above rises with height. The stability of the stratosphere results in a strongly layered structure in which thin layers of aerosol can persist for a long time. The small concentrations of water vapor mean that latent heat release or condensation becomes unimportant, so weather and clouds are rare. However, *mother-of-pearl clouds* are sometimes seen at altitudes of 20–30 km.

The stratosphere is a region of intense interactions among radiative, dynamical, and chemical processes, in which horizontal mixing of gaseous components proceeds much more rapidly than vertical mixing. The stratosphere is warmer than the upper troposphere, primarily because of a stratospheric ozone layer that absorbs solar ultraviolet radiation.

The chemical composition of the stratosphere is generally similar to that of the troposphere with some exceptions, the most notable of which are ozone and water vapor. The stratosphere is relatively dry. However, it is rich in ozone as it is the main region of ozone production. Ozone absorbs ultraviolet radiation from the Sun and with the low densities present at stratospheric altitudes, this absorption is an efficient mechanism of transferring kinetic energy to a relatively small number of molecules due to which the air temperature becomes high. Ozone in the upper stratosphere therefore acts as a heat source. Some of the heat is transferred down by subsidence and radiation, although the stratosphere as a whole remains warm at the top, where the temperatures are close to those at the Earth's surface, and cold at the bottom and therefore very stable.

The upper limit of the stratosphere is called the *stratopause*, which occurs at an altitude of 50–55 km, the level at which temperature ceases to increase with altitude.

1.4.3 Mesosphere

The mesosphere, a layer extending from 50 to 80 km, is characterized by decreasing temperatures with increasing altitude, reaching about 180 K at 80 km. Compared to lower regions, the concentrations of ozone and water vapor in the mesophere are negligible, hence the lower temperatures. Its chemical composition is fairly uniform. Pressures are very low. The *mesopause*, which separate the mesosphere from the next highest layer and like the other pauses, is a region where the temperature trend changes direction. Like the tropopause, however, the temperature of the mesopause can vary quite significantly, dropping as low as 150 K. In some instances *noctilucent clouds* can form here and affect global climate by absorbing and reflecting incoming solar radiation. The mesopause is the level at which the lowest atmospheric temperatures are usually found.

Middle atmosphere is the region lying between the troposphere and thermosphere which extends from approximately 10 to 100 km, comprising the stratosphere and mesosphere.

1.4.4 Thermosphere

The thermosphere is a region of high temperatures above the mesosphere. It includes the ionosphere and extends out to several hundred kilometers. The temperatures in this region are of the order of 500–2,000 K and the densities are very low. The thermosphere is that part of the heterosphere which does not have a constant chemical composition with increasing altitude. Rather, atoms tend to congregate in layers with the heavier species at lower altitudes.

The increase in temperature in the thermosphere is due to the absorption of intense solar radiation by the limited amount of molecular oxygen present. At an altitude of 100–200 km, the major atmospheric components are still nitrogen and oxygen, although at this high altitude gas molecules are widely separated. *Auroras* normally occur in this region between 80 and 160 km.

The *thermopause* is the level at which the temperature stops rising with height. Its height depends on the solar activity and is located between 250 and 500 km.

1.4.5 Exosphere

The exosphere is the most distant atmospheric region from Earth's surface. The upper boundary of the layer extends to heights of perhaps 960–1,000 km. The exosphere is a transitional zone between Earth's atmosphere and interplanetary space.

1.5 Structure of the Upper Atmosphere

The upper atmosphere is divided into regions based on the behavior and number of free electrons and other charged particles. The importance of the upper atmosphere is that instead of absorbing or reflecting radiation it deflects ionized particles.

1.5.1 Ionosphere

The ionosphere is the region of the Earth's atmosphere in which the number of ions, or electrically charged particles, is large enough to affect the propagation of radio waves. The ionosphere begins at an altitude of about 50 km but is most distinct

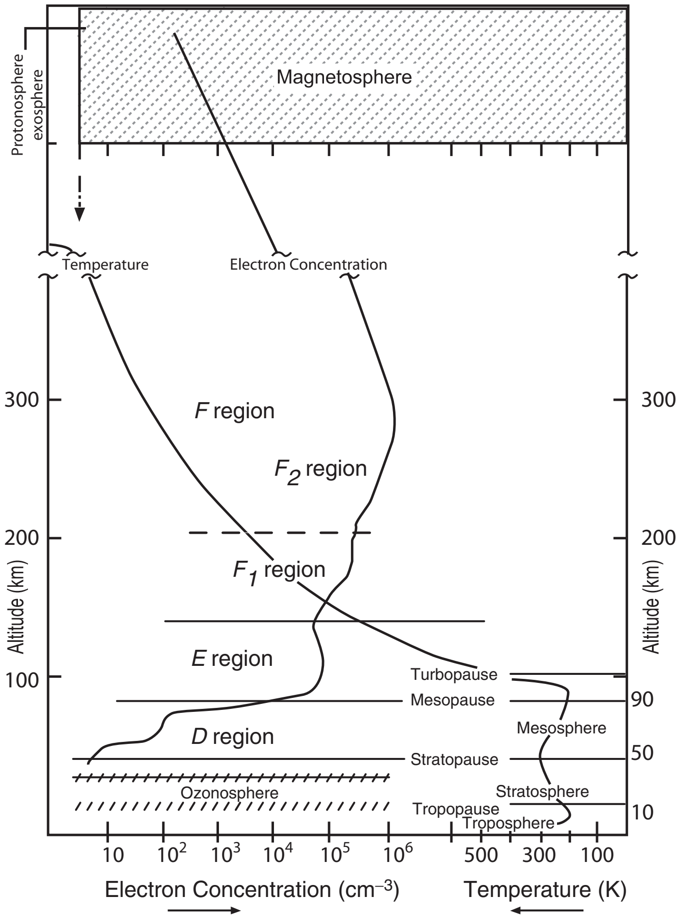

Fig. 1.5 A typical daytime profile of electron concentration of the ionosphere and the neutral atmosphere. A typical temperature profile in the reverse direction is also shown (Adapted from University of Leicester)

above about 80 km. The ionization is caused mainly by solar radiation at X-ray and ultraviolet wavelengths. The ionosphere is responsible for the long-distance propagation, by reflection, of radio signals in the shortwave and broadcast bands.

Vertical variation of the ionospheric layers in the Earth's atmosphere is depicted in Fig. 1.5. It can be seen that the ionosphere is highly structured in the vertical direction. It was first thought that discrete layers were involved, referred to as the D, E, F1, and F2 layers. However, the layers actually merge with one another to such an extent that they are now referred to as regions rather than layers. The very high temperatures in the Earth's upper atmosphere are colocated with the upper ionosphere since both are related to the effect of X-rays from the Sun. That is, the X-rays both ionize and heat the uppermost portion of the Earth's atmosphere. Tremendous variations occur in the ionosphere at high latitudes because of the dynamical effects of electrical forces and because of the additional sources of plasma production. The most notable is the visual aurora, one of the most spectacular natural sights.

The aurora has a poleward and equatorward limit during times of magnetic storms. Residents of the arctic regions of the northern hemisphere see the *Northern* lights in their southern sky. The aurora forms two rings around the poles of the Earth. The size of the rings waxes and wanes while wavelike disturbances propagate along its extent.

1.5.2 Plasmasphere

The plasmasphere is not really spherical but a doughnut-shaped region (a torus) with the hole aligned with Earth's magnetic axis. In this case the use of the suffix sphere is more in the figurative sense, like a *sphere of influence*. It is composed mostly of hydrogen ions (protons) and electrons, and is essentially an extension of the ionosphere. The torus has a very sharp outer edge called the *plasmapause*, which is usually some 4–6 Earth radii (19,000–32,000 km) above the equator. The inner edge of the plasmasphere is taken as the altitude at which protons replace oxygen as the dominant species in the ionospheric plasma, which usually occurs at about 1,000 km altitude.

1.5.3 Magnetosphere

The magnetic field of the Earth to a large extent shields it from the continual supersonic outflow of the Sun's ionized upper atmosphere, known as the solar wind. Inside of the plasmapause, geomagnetic field lines rotate with the Earth. Outside the plasmapause, however, magnetic field lines are unable to corotate because they are influenced strongly by electric fields produced by the solar wind. The magnetosphere is thus a nonspherical cavity in which the Earth's magnetic field is constrained by the solar wind and interplanetary magnetic field and shaped by the passage of the Earth. Consequently, the magnetosphere is shaped like an elongated teardrop with the tail pointing away from the Sun, as shown in the Fig. 1.6.

The outer boundary of the magnetosphere is called the *magnetopause* and marks the outer limit of Earth's gaseous envelope. The magnetopause is typically located at about 10 Earth radii (about 56,000 km) above the Earth's surface on the day side and stretches into a long tail, the *magnetotail*, a few million kilometers long (about 1,000 Earth radii) on the night side of the Earth. This is well past the orbit of the Moon (which is at 60 Earth radii), but the Moon itself is usually not within the magnetosphere except for a couple of days around the Full Moon.

Physical processes occurring within the magnetosphere modulate the energy flow carried by the solar wind to the Earth. But it is very responsive to changing solar wind conditions. At times the magnetosphere acts as a shield deflecting the incident

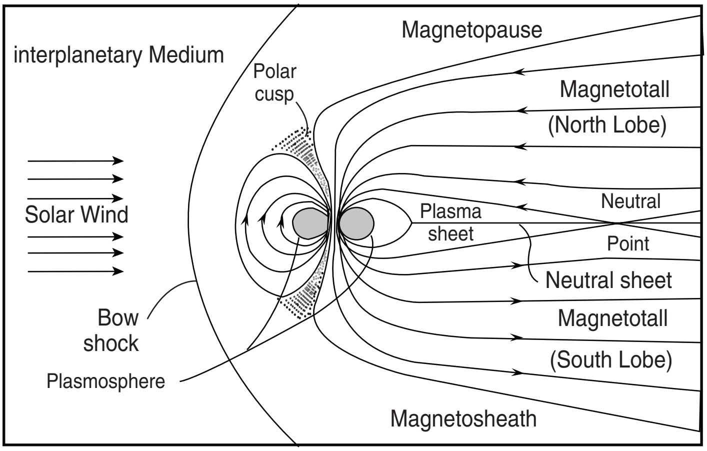

Fig. 1.6 Earth's magnetosphere with principal particle regions (C.T. Russel 1987, Courtesy: Terra Scientific Publishing Company, Tokyo)

energy; at other times it acts as an accelerator, driven by the solar wind, creating charged particle beams that hit the neutral upper atmosphere, causing it to light up with the brilliant forms of the polar aurora.

Since the scope of the book is on stratosphere–troposphere interactions, further discussion will be confined to the characteristics of Earth's lower and middle atmosphere.

1.6 The Tropopause

The tropopause is an important meteorological concept. It separates the troposphere from the stratosphere, i.e., two volumes of air with significantly different properties (Holton et al., 1995). In this region the air ceases to cool, and the air becomes almost completely dry. Basically, it is the boundary between the upper troposphere and the lower stratosphere that varies in altitude between the poles and the equator.

Tropopause is the transition layer between the troposphere and the stratosphere, where an abrupt change in temperature lapse rate usually occurs. It is defined by the World Meteorological Organization (WMO) as the lowest level at which the lapse rate decreases to $2\,\mathrm{K\ km^{-1}}$ or less, provided that the average lapse rate between this level and all higher levels within 2 km does not exceed $2\,\mathrm{K\ km^{-1}}$. Occasionally, a second tropopause may be found if the lapse rate above the first tropopause exceeds $3\,\mathrm{K\ km^{-1}}$ fall below this value. This so-defined thermal tropopause can be obtained from single temperature profiles and can be applied in both the tropics and the extratropics.

The tropopause is not a fixed boundary. Severe thunderstorms in the intertropical convergence zone (ITCZ) and over midlatitude continents in summer continuously push the tropopause upwards and as such deepen the troposphere. A pushing up of tropopause by 1 km reduces the tropopause temperature by about 10 K. Thus in areas and also in times when the tropopause is exceptionally high, the tropopause temperature becomes very low, sometimes below 190 K. Intense convective clouds in the tropics often overshoot the tropopause and penetrate into the lower stratosphere and undergo low-frequency vertical oscillations.

Tropopause height shows large variations with latitude, season, and even day-to-day. Latitudinal variation of the tropopause from the poles to the equator is schematically illustrated in Fig. 1.7. The tropopause height varies from 7–10 km in polar regions to 16–18 km in the tropics. Tropical tropopause is higher and colder, whereas polar tropopause is lower and warmer. The characteristic features of the tropopause over the tropics, midlatitude, and polar regions are illustrated in Table 1.2. The tropopause height also varies from troughs to ridges, with low tropopause height in cold troughs and high in warm ridges. Since these troughs and ridges propagate, the tropopause height exhibits frequent fluctuations at a particular location during midlatitude winters.

The highest tropopause is seen over south Asia during the summer monsoon season, where the tropopause occasionally peaks above 18 km. The oceanic warm pool of the western equatorial Pacific also exhibits higher tropopause height of 17.5 km. On the other hand, cold conditions lead to lower tropopause, evidently due to weak convection.

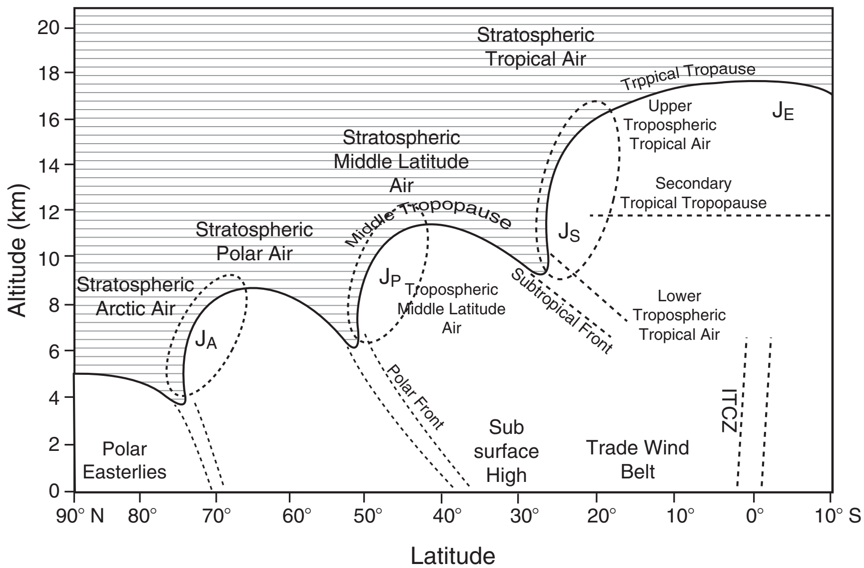

Fig. 1.7 Diagram showing the latitudinal variation of tropopause levels (Shapiro et al. 1987, Courtesy: American Meteorological Society)

Table 1.2 Characteristic features of tropopause at various latitude zones

Features	Tropical tropopause	Midlatitude tropopause	Polar tropopause
Location	Over tropics, between the two subtropical jet streams	Between polar and subtropical jet streams	North of polar jet
Height	$\sim$18 km	$\sim$12 km	6–9 km
Altitude	$\sim$80–100 hPa	$\sim$200 hPa	$\sim$300–400 hPa
Temperature	$\sim -80°$C	$\sim -60°$C	$\sim -45°$C
Potential temperature	$\sim$375–400 K	$\sim$325–340 K	$\sim$300–310 K
Character	Sharply defined, highest and coldest	Higher in summer and lower in winter	Often difficult to identify

1.6.1 Tropical Tropopause

The tropical tropopause layer (TTL) is the region of the tropical atmosphere that lies between the top of the main cumulus outflow layer ($\sim$12 km) and the thermal tropopause ($\sim$16 km). This layer is a transition layer between dynamical control of the vertical mass flux by tropospheric convection, and by the stratospheric circulation, and is crucial for understanding the dehydration of air entering the stratosphere. The annual cycle in transport of water vapor into the stratosphere is influenced by the seasonal variation of the stratospheric circulation and also by the annual cycle in TTL temperatures.

The coldest temperatures in the TTL occur over the equatorial west Pacific during northern hemisphere winter. Horizontal transport through this *cold trap* region causes air parcels that enter the TTL at other longitudes to be dehydrated to the very low saturation mixing ratios characteristic of the cold trap, and hence can explain why observed tropical stratospheric water vapor mixing ratios are often lower than the saturation mixing ratio at the mean tropopause temperature. Although the annual cycles in tropopause temperature and stratospheric pumping are the major controls on stratospheric water, the equatorial quasi-biennial oscillation (QBO) also plays a role in this regard.

Concisely, the TTL is defined as that region of the tropical atmosphere extending from the zero net radiative heating level (355 K, 150 hPa, 14 km) to the highest level that convection reaches (420–450 K, 70 hPa, 18–20 km). The TTL can be thought of as a transition layer between the troposphere and stratosphere and its structure and climatological aspects are very important for understanding the various coupling processes. Large-scale meteorological processes, like low and high pressure systems, can cause day-to-day variations in the tropopause height.

1.6.2 Tropopause Acronyms

The tropopause region exhibits a complex interplay between dynamics, transport, radiation, chemistry, and microphysics. This is particularly highlighted in the case of ozone and water vapor, which provide much of the climate sensitivity in this region. The tropopause region is therefore considered as a critical region for climate. Various acronyms used for the definition of tropopause are listed below (Haynes and Shepherd 2001).

Lapse-rate tropopause (LRT): LRT is the conventional meteorological definition of the tropopause, in both tropics and extratropics, as the base of a layer at least 2 km thick, in which the rate of decrease of temperature with height is less than $2\,\mathrm{K\,km^{-1}}$.

Cold-point tropopause (CPT): CPT is the level of minimum temperature. This is useful and significant in the tropics.

Tropical thermal tropopause (TTT): Since in the tropics the LRT and the CPT are usually less than 0.5 km apart (LRT being the lower) we ignore the distinction between them and refer to the TTT. The TTT is typically at 16–17 km.

Secondary tropical tropopause (STT): STT is the level of maximum convective outflow, above which the lapse rate departs from the moist adiabat. The STT is typically at 11–12 km.

Clear-sky radiative tropopause (CSRT): CSRT is the level at which the clear-sky heating is zero. Below the CSRT there is descent on average (outside convective clouds). Above the CSRT there is ascent on average. The CSRT is typically at 14–16 km.

It is believed that the tropical tropopause is a source region of tropospheric moist air entering the stratosphere. Observational studies (Gettelman and Forster 2002) indicate that most of the tropical deep convection does not reach the cold point tropopause but ceases a few hundred meters lower, sandwiching the TTL, i.e., a relatively undisturbed body of air, subject to prolonged chemical processing of air parcels which may slowly ascend into the stratosphere.

1.6.3 Dynamic Tropopause

More recently the so-called *dynamical tropopause* has become popular. Dynamic tropopause is used with potential vorticity instead of vertical temperature gradient as the defining variable. There is no universally used threshold, but the most common ones are that the tropopause lies at the 2 PVU (potential vorticity unit) or 1.5 PVU surface. This threshold will be taken as a positive or negative value (e.g., 2 and −2 PVU), giving surfaces located in the northern and southern hemisphere respectively. To define a global tropopause in this way, the two surfaces arising from the positive and negative thresholds need to be joined near the equator using another type of surface such as a constant potential temperature surface as illustrated in Fig. 1.8.

A PV anomaly is produced by the intrusion of stratospheric air into the upper troposphere. An upper level PV anomaly advected down to middle troposphere is

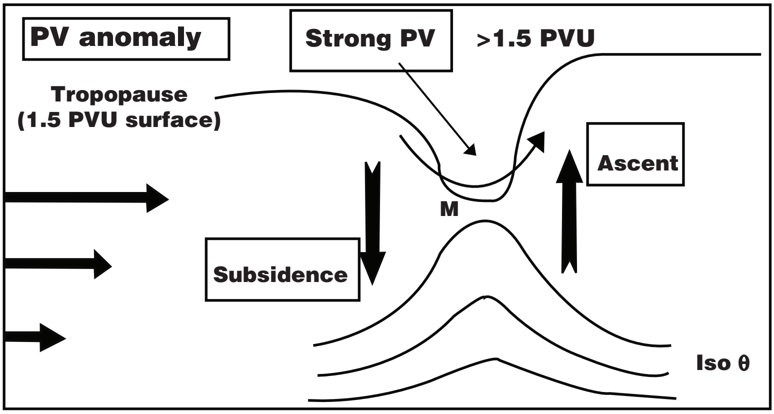

Fig. 1.8 A typical dynamical tropopause (Adapted from EUMeTrain)

called tropopause dynamic anomaly or folding of the dynamical tropopause. Due to PV conservation, the anomaly leads to deformations in vertical distribution of potential temperature and vorticity. In a baroclinic flow increasing with height, the intrusion of PV anomaly in the troposphere produces a vertical motion. The deformation of the isentropes imposes ascending motion ahead of the anomaly and subsiding motion behind it.

For conservative flow, the dynamical tropopause (unlike the thermal tropopause) is a material surface; this is an advantage for instance when considering the exchange of mass across the tropopause (Wirth 2003; Wirth and Szabo 2007). Despite overall similarities between the thermal and dynamical tropopause, they are certainly not identical and in specific situations there may be significant differences. For an atmosphere at rest *potential vorticity (PV)* is essentially a measure of static stability, and one can basically enforce both tropopauses to be at the same altitude through a suitable choice of the PV value for the dynamical tropopause.

1.6.4 Ozone Tropopause

Apart from *thermal* and *dynamical* tropopauses, there is another category for the definition of the tropopause based on the ozone content (Bethan et al. 1996) called *ozone tropopause*. In most seasons, ozone-mixing ratio similar to PV features a sharp positive vertical gradient at a particular altitude somewhere in the tropopause region. It can be used for defining an ozone tropopause from a single ozone sounding. In addition, ozone-mixing ratio like PV is approximately materially conserved on synoptic timescales. Therefore one may expect that the ozone tropopause would behave like the dynamical tropopause.

1.6.5 Tropopause Folds

Tropopause fold is the extrusion of stratospheric air within an upper-tropospheric baroclinic zone which slopes downward from a normal tropopause level (∼200–300 hPa) to the middle troposphere (∼500–700 hPa). The tropopause fold is a mesoscale feature which forms in response to strong descent at the tropopause level. It constitutes an intense phase of upper tropospheric frontal development in which the tropopause undulation collapses. In regions of confluent flow the tropopause may be deformed such that it will form a fold (as shown in Fig. 1.9) which will decay after 1 or 2 days. During the buildup phase of a fold the flow is generally conservative, whereas the decay phase is characterized through nonconservative flow, e.g., diabatic heating and turbulent mixing. It is these nonconservative processes which achieve the stratosphere–troposphere exchange.

The most vigorous tropopause folds occur during the winter and spring and are less frequent than cyclone development. They are usually observed downstream from a ridge, where there is large-scale descent in the entrance region of an upper-level jet streak. Ozone-rich air originates in the lower stratosphere, west of the trough axis on the cyclonic side of the upper-level jet streak. This airstream then descends anticyclonically to the lower troposphere east of the surface high-pressure system or crosses the trough axis and ascends.

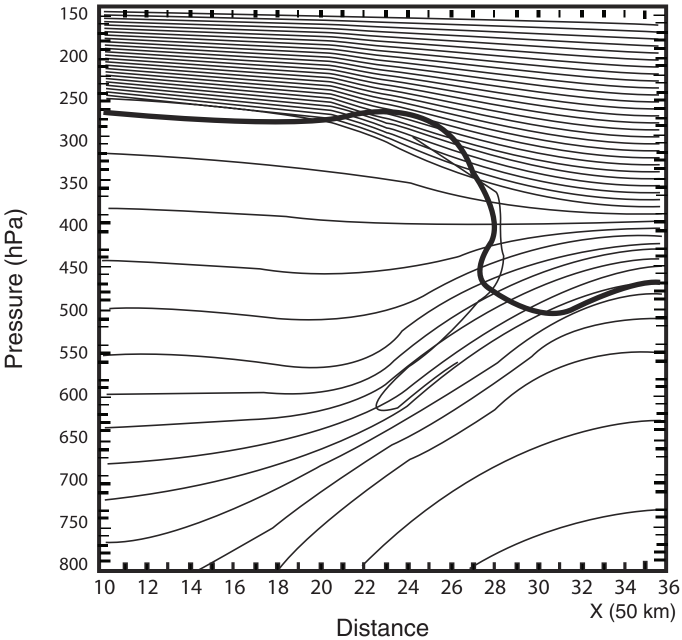

Fig. 1.9 Cross section of an idealized tropopause fold (Adapted from: G. Hartjenstein 1999)

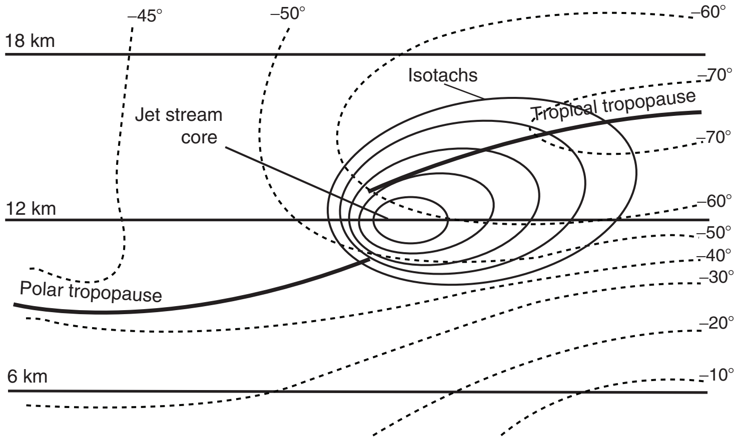

Fig. 1.10 Discontinuity in the tropical and polar tropopauses and the formation of jetstreams (Adapted from The Green Lane, Environmental Canada)

Two major types of tropopause folds are noted. One is associated with the polar front jet (PFJ), which may extend deep into the troposphere along the polar front. In some PFJ tropopause folds, significant intrusion of stratospheric air deep into the troposphere occurs. The other one is associated with the subtropical jet stream (STJ) and subtropical front, which is confined in the upper troposphere only and rarely extends downward below 500 hPa. The positions of the tropical tropopause and the polar tropopause along with the formation of subtropical jet stream over the midlatitude region are schematically shown in Fig. 1.10.

Vertical intrusions of the dynamical tropopause into the troposphere, which are folded due to differential isentropic advection, are also known as tropopause folds.

1.6.6 Importance of Tropopause to Tropospheric Weather Events

Tropopause locates the base of the stratosphere which is a layer of high static stability. Thus it acts to damp vertical motion on both large and small scales. It is most noticeable when it acts as an upper lid to deep convection causing the spreading out of cirrus anvil clouds. Tropopause marks the reversal of the tropospheric north–south temperature gradient. This causes the jet stream to be near the tropopause, via the thermal wind concept (Asnani 2005).

Tropopause folds may play an important role in cyclogenesis due to the conservation of potential vorticity principle. The potential vorticity is quite large in the stratosphere. If a body of air with large potential vorticity enters the troposphere through a tropopause fold and encounters a region of lower static stability, then relative vorticity must increase to conserve the potential vorticity, which affects the circulation pattern of the troposphere.

1.7 Climatology of the Lower and Middle Atmosphere

The vertical distribution of temperature and wind structure of the lower and middle atmosphere has been studied extensively for several decades using a variety of techniques. A large number of measurements have been made by using balloons, aircrafts, radiosonde, rocket and satellite observations, both spatially and temporally, on a global scale. Based on the global observations, and modeling information, Stratospheric Process and Its Influence on Climate (SPARC) has prepared a reference climatology of the middle atmosphere (SPARC 2002). In this section, the temperature and wind climatology of the troposphere and stratosphere are mainly discussed.

1.7.1 Temperature

Figure 1.11 illustrates the zonal mean temperature extent from surface to 90 km and from south pole to north pole for January derived from METEO analyses (1,000–1.5 hPa), and a combination of HALOE plus MLS data above 1.5 hPa (Randel et al. 2004). The thick dashed lines in the figure denote the mean tropopause (taken from NCEP reanalyses) and mean stratopause, obtained by the local temperature maximum near 50 km.

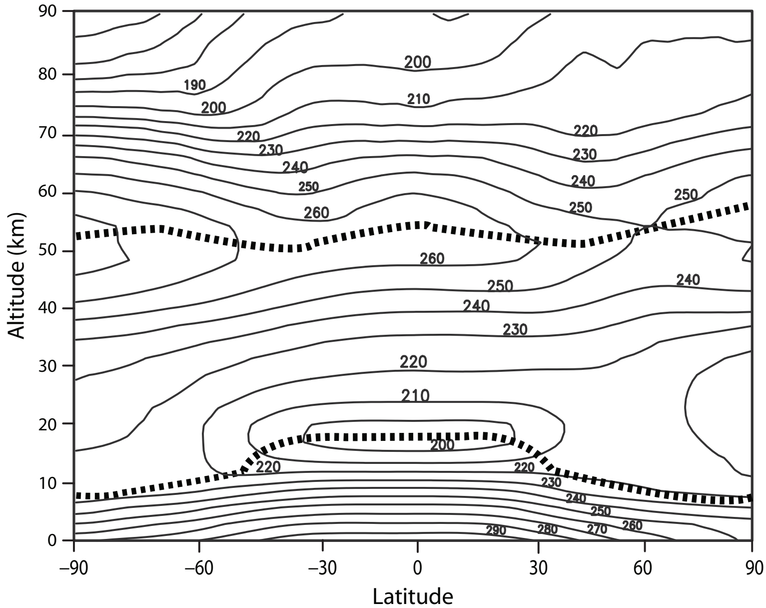

Fig. 1.11 Climatological zonal mean temperature for January Courtesy (Randel et al. 2004; Courtesy SPARC)

The temporal and latitudinal variability of mean temperature profile shows that there is considerable latitudinal and seasonal variability. Temperature decreases with latitude in the troposphere. The latitudinal gradient is about twice as steep in the winter hemisphere compared to that in the summer hemisphere. The tropopause is much higher and colder over the tropics than over the polar regions.

The latitudinal distribution of temperature in the lower stratosphere is rather complicated. The summer hemisphere has a cold equator and a warm pole. The winter hemisphere is cold at both equator and pole with a warmer region in middle latitudes. The cold pool of stratospheric air over the winter pole is highly variable. On occasions, it disappears for a period of a few weeks during midwinter. During these so-called *sudden stratospheric warmings*, the stratospheric temperatures over individual stations have been observed to rise by as much as 70°C in 1 week (Labitzke and van Loon 1999).

At the stratopause, there is a monotonic temperature gradient between the warm summer pole and the cold winter pole. At the mesopause, the situation is exactly opposite: the summer pole is cold, the winter pole is warm. Temperature has a pronounced diurnal variability in certain regions of the atmosphere. The strongest variability is observed in the upper thermosphere. In this region, day to night temperature differences are on the order of several hundred degrees.

There are also significant, but much smaller, diurnal variations around the stratopause level. These give rise to strong tidal motions in the Earth's upper atmosphere. The tidal motions manifest as regular oscillations in surface pressure which are prominent in the tropics. In the middle and upper troposphere, the day to night changes are typically less than a degree. However, in the lowest few kilometers over land the changes are somewhat larger. At the Earth's surface, especially over land, the diurnal range is typically on the order of 10°C. It may even exceed 20°C over high altitude desert regions.

1.7.2 Wind

The large-scale features of the atmospheric zonal wind circulation from surface to 90 km are shown in Fig. 1.12. This is a latitude–height cross section of the zonal wind component, averaged with respect to longitude, during the month of January derived from METEO analyses (1,000–1.5 hPa), and a combination of HALOE plus MLS data above 1.5 hPa, similar to that of temperature (Randel et al. 2004). The tropopause and stratopause levels are marked in dashed lines in the figure.

The mean zonal flow in the winter hemisphere equatorward of 40° latitude is similar, with stronger westerlies about 40 m s^{-1} at the 200 hPa level. The maximum wind in the southern hemisphere (SH) is about 2–3° latitude nearer the equator and is about 5 m s^{-1} weaker than the northern hemisphere (NH) winter maximum. Poleward of 40°S latitude, the zonal winds differ appreciably in winter, with stronger winds in the SH. A westerly maximum in the upper troposphere that continues into

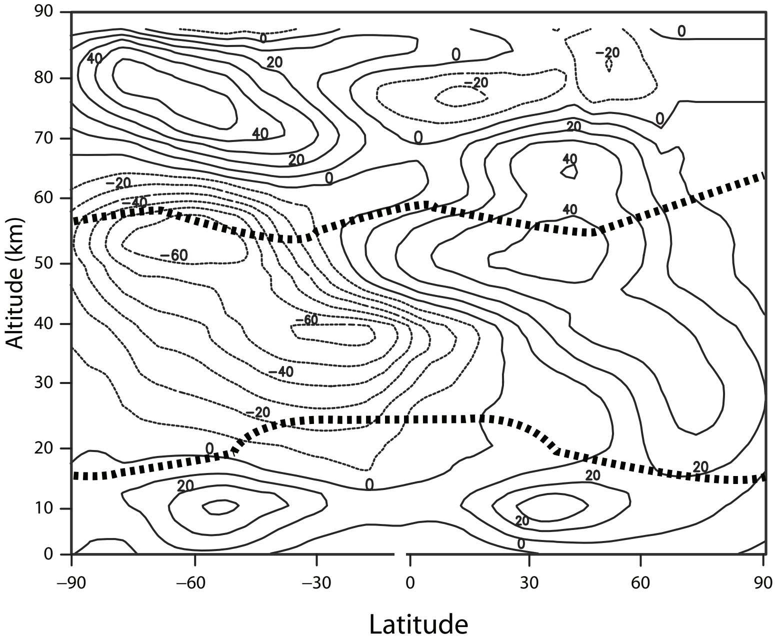

Fig. 1.12 Climatological zonal mean temperature for January (Randel et al. 2004, Courtesy: SPARC)

the stratosphere is evident between 50° and 60°S in accordance with the upward-increasing meridional temperature contrast poleward of 45°S. The distribution of wind differs considerably between the summer hemispheres. The upper troposphere westerly maximum is nearly twice as strong in the SH and is located farther poleward than the peak in the NH. In the middle and upper troposphere the tropical easterlies are much stronger in the NH than in the SH, and in the subtropics the westerlies are much stronger in the SH.

Prominent features are cores of strong westerly winds in middle latitudes at an altitude of 10 km. However, the strongest zonal winds occur in the mesosphere at an elevation of 60 km. Again there are two jet cores in middle latitudes, the stronger in the winter hemispheres westerly; the other in the summer hemisphere is easterly. During the equinoxes, these jets undergo dramatic reversals as the latitudinal temperature gradient reverses. Certain important features of the longitudinally averaged zonal wind field do not show up explicitly in Fig. 1.12. For example, the *sudden stratospheric warming* phenomenon is accompanied by large changes in the longitudinally averaged zonal wind at high latitudes in the winter stratosphere. The midwinter warmings are accompanied by a pronounced weakening of the westerlies at stratospheric levels. Sometimes the westerlies disappear altogether. These changes in the stratosphere have little effect on the wind structure in the troposphere.

1.7.2.1 Zonal Mean Winds in the Equatorial Stratosphere

The direction and speed of the zonal mean winds in the tropics are dominated by the equatorial lower stratospheric phenomenon known as *Quasi Biennial Oscillation* (QBO). QBO in equatorial zonal winds alter from easterly to westerly at high altitudes above 30 km, with the easterly and westerly phases descending downward with height, so that the easterly winds will at one point be above the westerly winds and the westerly winds will at one point be above the easterly winds. The QBO extends to aproximately 10–15° latitudes on either side of the equator, although the effects of the QBO are felt in the subtropics to 30° latitude and even in high latitude regions.

The average period of oscillation of QBO is about 28 months, although the QBO period varies from about 20 to 30 months. The strongest easterly winds are about 30 $\mathrm{m\,s^{-1}}$, while the strongest westerly winds are typically 20 $\mathrm{m\,s^{-1}}$. Westerlies typically begin to descend in June through August, although this is not a rigid rule. The QBO phenomenon is discussed in detail in section 1.7.8 and also in the later chapters of this book.

1.7.2.2 Zonal Mean Winds in the Midlatitude Stratosphere

The zonal mean winds in the midlatitude winter hemisphere are westerly, with a maximum velocity of 80 $\mathrm{m\,s^{-1}}$ at 65 km altitude and 40° latitude. In summer hemisphere it becomes easterly, with a maximum velocity of about 50 $\mathrm{m\,s^{-1}}$ at 65 km altitude and 40° latitude.

In the northern hemisphere, zonal mean winds change from westerly to easterly in May, starting at the highest latitudes and altitudes and moving downward towards the tropics. Winds change from easterly to westerly in September, once again starting from high latitudes and altitudes.

As in the troposphere, the actual winds in the stratosphere have significant meridional components. Air parcels in midlatitudes typically traverse the globe in 1–2 weeks, depending on the location and the circumstances.

1.7.3 Diurnal Cycle

Time variations of the atmospheric state have two periodic components known as a *diurnal cycle* and an *annual cycle*, which are responses to the periodic variations of the external forcings due to the Earth's rotation and revolution about the Sun, respectively. Examples of the former are land and sea breeze, mountain and valley winds, and thermal tides, which are periodic responses of the atmosphere to the diurnal differential heating by the Sun on local or global scales. A monsoon climate with dry and rainy seasons is a clear example of the annual cycle.

1.7.4 Annual Oscillation

The *Annual Oscillation (AO)* is defined as the tendency of the lower stratospheric winds to become easterlies in the summer hemisphere and westerlies in the winter hemisphere. The AO is primarily an extratropical phenomenon and does not interact strongly with tropical circulation systems.

The seasonal march and the year-to-year variation of the stratospheric circulation are significantly different between the northern and southern hemispheres. In the northern hemisphere, monthly mean (steady state) planetary waves show the maximum amplitudes in midwinter associated with the occurrence of sudden warmings, but in general the phase of east–west wave number 1 is almost fixed due to the topographic effect of the surface.

On the other hand, planetary waves in the southern hemisphere stratosphere show large amplitudes in late winter or early spring (September and October), and the off-pole pattern of the polar vortex, which means the phase and amplitude of wave number 1 varies from year to year, as can be seen in the Antarctic ozone hole. The generation and maintenance of SH stratospheric waves are quite sensitive to the transient wave activity in the troposphere (Hirota and Yasuko 2000).

1.7.5 Semiannual Oscillation

One of the significant features of the low latitude upper stratosphere is the semiannual oscillation (SAO) of zonal winds. It is reported that the SAO is not generated by the solar declination changes in the low latitudes during the course of the year. It is however known that changes in the eddy momentum deposited to the zonal winds in the upper stratosphere and lower mesosphere is responsible for the equatorial westward current phase of the SAO which occurs shortly after the equinoxes.

Vertical variation of the amplitudes of QBO and SAO in temperature in the troposphere and stratosphere over the equatorial region derived from the ERA40 reanalysis data are shown in Fig. 1.13. In temperature, the QBO and SAO are generally weak in their amplitude but have the same phase in the troposphere. The amplitude of QBO increases steeply in the lower stratosphere and it peaks around 25 km. Above this level, the amplitude of QBO decreases with height. In the case of SAO, the amplitude increases steadily from the lower stratosphere and reaches the maximum around 40 km. At this level the amplitude of SAO in temperature is higher than the peak amplitude of QBO at 25 km. SAO in temperature decreases at higher levels.

Figure 1.14 gives the zonal wind amplitudes of QBO and SAO with height in the troposphere and stratosphere over the equator. In zonal wind also, amplitudes of QBO and SAO are weak and in the same phase in the entire troposphere. The QBO increases in amplitude rapidly with height in the lower stratosphere and reaches the maximum around 30 km altitude, and decreases in the upper stratosphere. On the other hand, the amplitude of SAO remains weak in the lower and middle

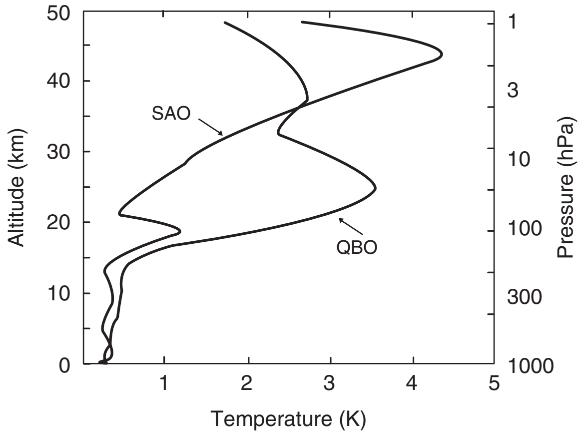

Fig. 1.13 Vertical variations in the amplitudes of SAO and QBO in temperature (Courtesy: W.J. Randel)

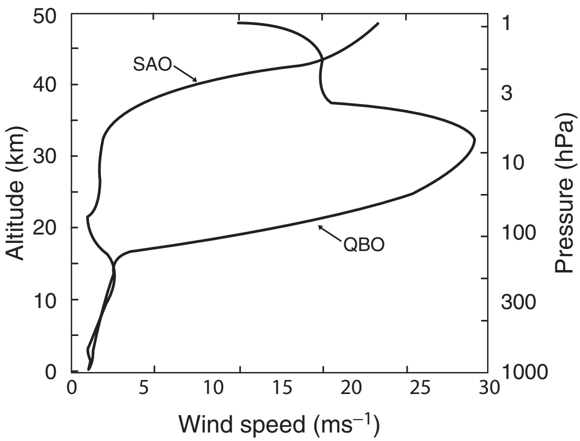

Fig. 1.14 Vertical distribution of the amplitudes of SAO and QBO in zonal wind over the equator (Courtesy: W.J. Randel)

stratosphere, but increases rapidly with height in the upper stratosphere. In this region, the amplitudes of the QBO and SAO are out of phase.

The semiannual oscillation is confined to tropics, with its phase decreasing with altitude in a manner similar to QBO. These features strongly suggest that the westerly phase of SAO should be due to the westerly momentum deposition at the relevant altitudes by Kelvin waves. The very rapid dissipation with heights of the long-period Kelvin waves producing the westerly phase of the QBO implies that the longer-period Kelvin modes would not be available to transport energy and momentum to the mesospheric levels. However, the shorter-period Kelvin waves are not absorbed significantly in the lower stratosphere and these waves with very small amplitudes carry large enough westerly momentum for the westerly phase of the SAO. So far there does not exist a satisfactory mechanistic model for the SAO in the upper stratosphere and lower mesosphere.

1.7.6 Interannual and Intraseasonal Oscillations

Interannual variation is a year-to-year variation which is defined as a deviation from the climatological annual cycle of a meteorological quantity. It can be caused by a variation of an external forcing of the atmospheric circulation system, or can be generated internally within the system. On the other hand, *intraseasonal variation* is a low-frequency variation within a season, and it is considered to be the result of internal processes which may exist even under constant external conditions.

Intraseasonal and interannual variations are defined as deviations from the periodic annual response. In general, intraseasonal variation means low-frequency variation with week-to-week or month-to-month timescales, while interannual variation means year-to-year variation. Some part of these variations is a response to the time variations of the external forcings or boundary conditions of the atmospheric circulation system, while the rest is generated internally within the system.

The time variations of the troposphere and the stratosphere are to be considered independently, partly because the heat capacity is entirely different between the layers and the adjacent ocean and land. In recent years, however, interactions between the troposphere and the stratosphere in the intraseasonal and interannual timescales have drawn attention to a possible stratospheric role in climate.

1.7.7 Jet Streams

Among the fascinating features of upper-air circulations are discontinuous bands of relatively strong winds of more than 30 m s^{-1}, called jet streams. It is a narrow band of air that moves around the globe at relatively very high speed near the tropopause height. Strong vertical shear is experienced in the vicinity of the jet streams. As with other wind fields that increase with increasing height, jet streams can be explained as an application of the thermal wind equation. They are located above areas of particularly strong temperature gradients (e.g., frontal zones). In such areas, the pressure gradients and the resulting wind speeds increase with increasing height so long as the temperature gradients persist in the same direction. In general, this will extend to the tropopause, after which the temperature gradient reverses direction and the wind speeds diminish. Thus, jet streams are usually found in the upper troposphere (i.e., at levels of 9–18 km). The positions of jet streams in the atmosphere are illustrated in Fig. 1.15.

Because regions of strong temperature gradients can be created in different ways, there are several classes of jet streams. Perhaps the most familiar is the polar-front jet

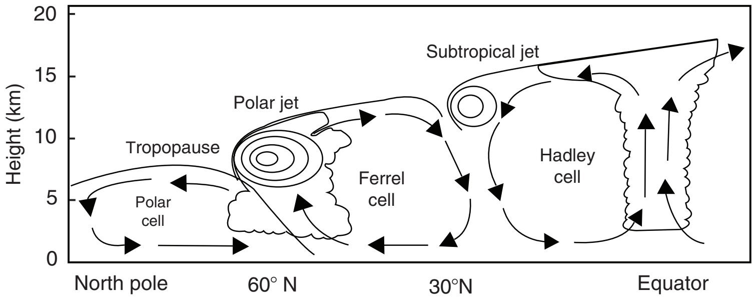

Fig. 1.15 Positions of jet streams in the meridional circulation (adapted from National Weather Service, NOAA)

stream. As noted earlier, the polar front is the boundary between polar and midlatitude air. In winter this boundary may extend equatorward to 30°, while in summer it retreats to 50–60°. Winter fronts are also distinguished by stronger temperature contrasts than summer fronts. Thus, jet streams are located more equatorward in winter and are more intense during that time with jet core wind speed exceeding 75 $\mathrm{m\,s^{-1}}$.

A second jet stream is located at the poleward limit of the equatorial tropical air above the transition zone between tropical and midlatitude air. This *subtropical jet stream (STJ)* is usually found at latitudes of 30–40° in general westerly flow. This jet may not be marked by pronounced surface temperature contrasts but rather by relatively strong temperature gradients in the mid-troposphere. Moreover, when the *polar-front jet (PFJ)* penetrates to subtropical latitudes, it may merge with the subtropical jet to form a single band.

Also peculiar to the tropical latitudes of the northern hemisphere is a high-level jet called the *tropical easterly jet stream (TEJ)*. Such jets are located about 15° N over continental regions due to the latitudinal heating contrasts over tropical landmasses that are not found over the tropical oceans.

The location of all three jet streams in relation to other mean meridional circulation features is shown in the above figure. Jet streams occur in both hemispheres. Those in the southern hemisphere resemble the northern hemispheric systems, though they exhibit less day-to-day variability due to the presence of smaller landmasses.

1.7.8 Quasi-Biennial Oscillation (QBO)

The QBO was discovered in the 1950s, but its origin remained unclear for some time. Rawinsonde soundings showed that its phase was not related to the annual cycle, as is the case for all other stratospheric circulation patterns. In the 1970s it was recognized that the periodic wind change was driven by atmospheric waves emanating from the tropical troposphere that travel upwards and are dissipated in the stratosphere by radiative cooling.

As mentioned in section 1.7.2.1, QBO is an east–west oscillation in stratospheric winds characterized by an irregular period averaging 28 months. In the equatorial region where the QBO is dominant, easterlies are typically 30 m s^{-1} and westerlies near 20 m s^{-1}. The QBO is seen between 100 hPa and 2 hPa, with maximum amplitude near 10 hPa (Hamilton et al. 2004). Figure 1.16 is the update of the time–height cross section based on radiosonde observations from stations near the equator since 1953 (Marquardt and Naujokat 1997). The equatorial winds are dominated by alternating easterly and westerly wind regimes, with a period varying from 22 to 36 months. These wind regimes propagate irregularly downward, with easterly shear zones tending to propagate more slowly and less irregularly. The QBO may also be seen in temperature, and it dominates the interannual variability of total ozone in the tropics.

The amplitude of the QBO decreases rapidly away from the equator. However, observations and theory show that the QBO affects a much larger region of the atmosphere. Through wave coupling, the QBO affects the extratropical stratosphere during the winter season, especially in the northern hemisphere where planetary wave amplitudes are large. These effects also appear in constituents such as ozone. In the high-latitude Northern winter, the QBO's modulation of the polar vortex may affect the troposphere through downward penetration. Tropical tropospheric observations show intriguing quasi-biennial signals which may be related to the stratospheric QBO (Baldwin et al. 2001). The QBO has been even linked to variability in the upper stratosphere, mesosphere, and ionospheric F layer.

1.7.9 Mean Meridional Winds

Schematic representation of meridional circulation is illustrated in Fig. 1.17. The arrows indicate the direction of air movement in the Hadley cell, Ferrel cell, and the polar cell in the meridional plane. The position and the direction of flow of the tropical easterly jet stream, subtropical jet stream, and the polar jet stream are also depicted.

The Hadley cell ends at about 30° north and south of the equator because it becomes dynamically unstable, creating eddies that are the reason for the weather disturbances of the midlatitude belts. These eddies force a downward motion just south of the jet axis and an upward motion between 40° and 60° north and south of the equator, forming the Ferrel cell. The eddies are also responsible for spreading the westerlies down to the surface.

1.7.10 Zonally Averaged Mass Circulation

The annually averaged atmospheric mass circulation in the meridional plane is depicted in Fig. 1.18. The arrows indicate the direction of air movement in the longitudinal plane. The total amount of mass circulating around each cell is given by the largest value in that cell, based on the NCEP-NCAR reanalysis data for the period 1958–2007.

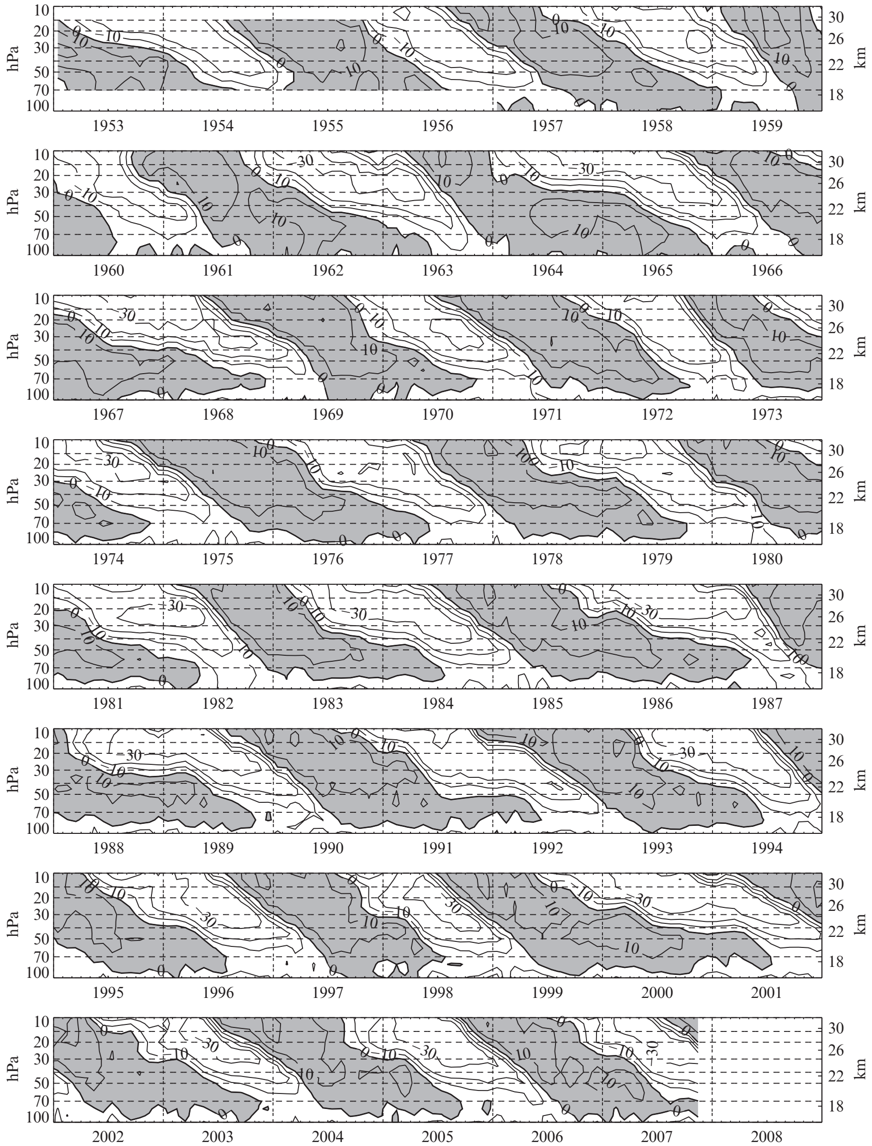

Fig. 1.16 Time–height section of zonal wind at equatorial stations showing the QBO. Isopleths are in meters per second. The data are from Canton Island (January 1953 to August 1967), Canton Island (September 1967 to December 1975), Gan/Maldive Islands (January 1976 to April 1985), and Singapore (May 1985 to August 1997) and updated upto 2007 November (Courtesy: Marquadt and Naujokat, Stratospheric Research Group, FSU Berlin)

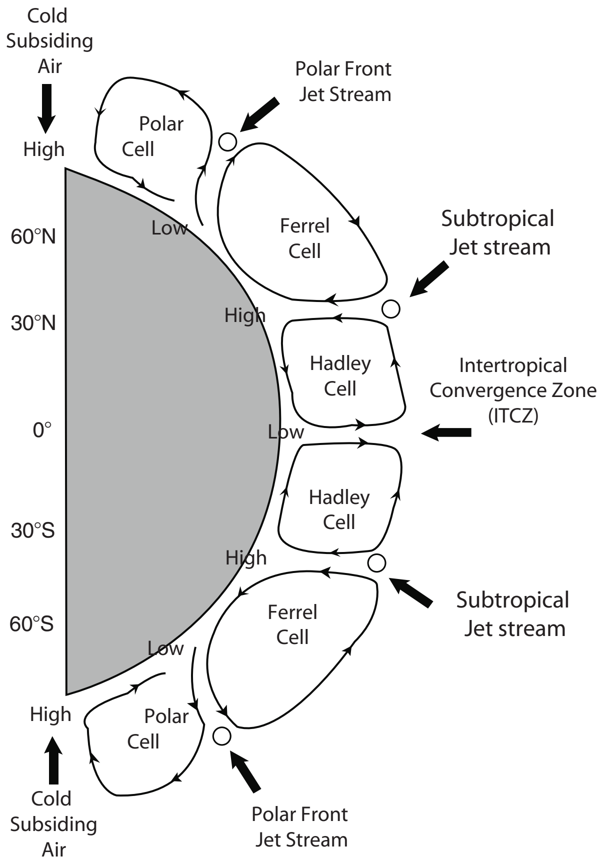

Fig. 1.17 Mean meridional mass circulation in the atmosphere (Adapted from RMIT University)

The rising motion in the tropics is capped from above by the stratosphere, where the air warms with height, thus suppressing upward motion. The law of mass continuity requires the air to move away from the tropics, northward and southward as in the diagram. This motion amounts to an upper-level mass divergence, forced by the rising motion. Again, for reasons of mass continuity, the diverging upper-level tropical air must return to the surface poleward of the equator. At the same time, mass continuity at the surface requires low-level convergence and the movement of air toward the equator (Andrews et al. 1987; McPhaden et al. 1998). A typical pattern of mean meridional circulation and surface winds during the boreal summer season is illustrated in Fig. 1.19.

The meridional winds are much weaker than the zonal winds. The typical transit time for air parcels to travel from the tropics to the pole is many months in the lower stratosphere, and longer at higher altitudes because the tropospheric air enters the stratosphere through the tropical tropopause and leaves the stratosphere at high latitudes near and above 60° latitude.

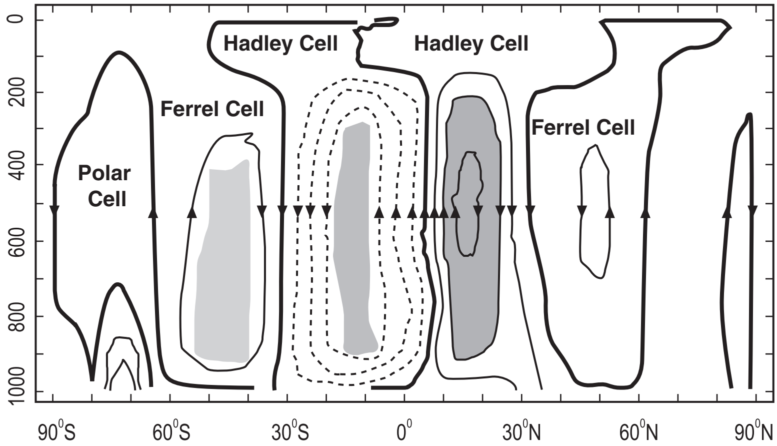

Fig. 1.18 The zonally averaged mass circulation. The arrows depict the direction of air movement in the meridional plane. (Source: NCEP/NCAR Reanalysis data)

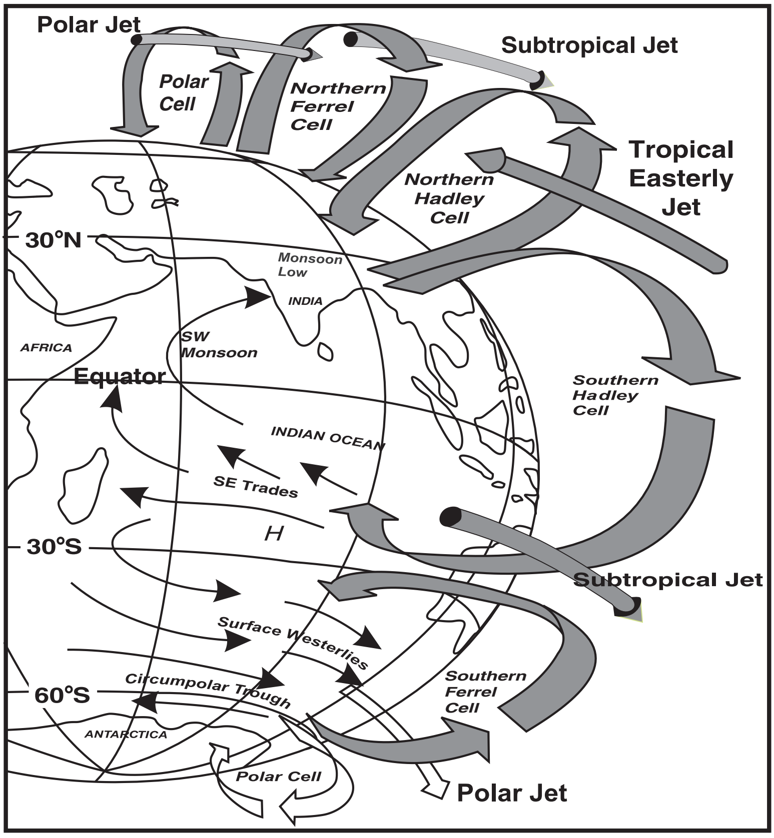

Fig. 1.19 Illustration of mean meridional mass circulation (broader shaded arrows), jetstreams (thick shaded arrows) and surface winds (thin arrows) during the northern hemispheric summer (Adapted from RMIT University)

The mean zonal stratospheric wind is roughly 20 m s^{-1}. That means the zonal wind in the stratosphere travels at a speed of 1,700 km day^{-1} and it takes about 10 days to circle the globe. On the other hand, the mean meridional stratospheric wind is relatively very weak. The mean meridional stratospheric wind is approximately 0.1 m s^{-1} and it takes nearly 3 years to travel 6,000 km. The mean vertical wind in the lower stratosphere is 2×10^{-4} m s^{-1}, which shows that the vertical wind in the stratosphere moves at a speed of 5 km year^{-1}.

1.7.11 The Polar Vortex

During the winter polar night, sunlight does not reach the south pole. A strong circumpolar wind develops in the middle to lower stratosphere. These strong winds constitute the *polar vortex* (see Fig. 1.20). Wind speeds around the vortex may reach 100 m s^{-1}. The vortex establishes itself in the middle to lower stratosphere. It is important because it isolates the very cold air within it. In the absence of sunlight the air within the polar vortex becomes very cold. Special clouds can therefore form once the air temperature gets to below $-80°$C.

Near the Arctic circle the situation is considered to be less severe because its polar vortex is not as well defined as that of the Antarctic; thus, the Arctic stratosphere is warmer than its Antarctic counterpart.

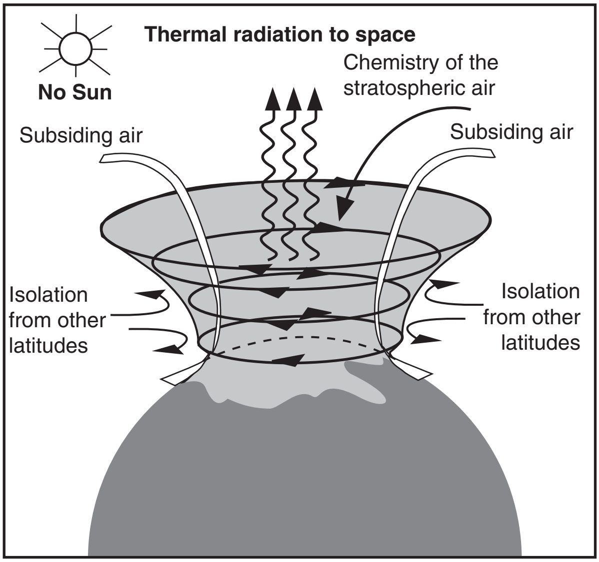

Fig. 1.20 Schematic representation of Polar Vortex (Adapted from J. Hays and P. deMenocal, University of Columbia)

The stratosphere is cooled by the greenhouse gases. The extent of the cooling is more pronounced in the higher levels of the stratosphere. The differential cooling is suspected to have contributed to the tightening of the vortex. Contraction of the polar vortex will affect the weather changes throughout the northern hemisphere (Kodera and Kuroda 2000). Prior to 1970 the polar vortex was volatile in nature. Strengthening and weakening of the polar vortex occurred from week to week or month to month, especially during winter. After the 1970s, the vortex has shown a considerable preference toward strengthening.

The eastward circulation of the Antarctic polar vortex is strongest in the upper stratosphere and strengthens over the course of the winter. The polar night jet is important because it blocks the transport between the southern polar region and the southern midlatitudes. It acts as a barrier and effectively blocks any mixing of air between inside and outside the vortex during the winter (Baldwin et al. 2003). Thus the ozone-rich air in the midlatitudes cannot be transported into the polar region.

Isolation of polar air allows the ozone loss processes to proceed without hindrance and replenishment by intrusions of ozone-rich air from midlatitudes. This isolation of the polar vortex is a key ingredient to polar ozone loss, since the vortex region can evolve without being disturbed by the more conventional chemistry of the midlatitudes. The polar night jet over the Arctic is not as effective in keeping out intrusions of warmer, ozone-rich midlatitude air. This is because there is more wave activity and hence more north–south mixing of air in the northern hemisphere than in the southern hemisphere.

1.8 Other Major Events in Stratosphere–Troposphere Interactions

In addition to the regular changes in the circulation and thermal structure of the stratosphere, there are several spectacular events taking place in the stratosphere which may often influence the tropospheric characteristics. Some of these events extend to larger areas, sometimes on a global scale, and affect tropospheric weather systems. In the following section, some of such major events are presented.

1.8.1 Polar Stratospheric Clouds

Polar stratospheric clouds (PSCs) are important components of the ozone depletion process in the polar regions of the Antarctic and the Arctic. PSCs provide the surfaces upon which chemical reactions involved in ozone destruction take place. These reactions lead to the production of free chlorine and bromine, released from CFCs and other ozone-depleting chemicals (ODCs) which directly destroy ozone molecules.

Since the stratosphere is very dry, the clouds formed in this region entirely differ in character to those formed in the moist troposphere. In the extreme cold condition

Fig. 1.21 Polar stratospheric clouds over Kiruna, Sweden (photo by H. Berg, Forschungszebntrum, Karlsrube)

of the polar winter, two types of PSCs may form. Type I clouds, consisting of mainly nitric acid and sulphuric acid, are more frequent. Whereas, Type II clouds are rarer and contain water or ice which form only below $-90°C$. Photograph of a polar stratospheric cloud is shown in Fig. 1.21.

Since these clouds are located at high altitudes and due to the curvature of the surface of the Earth, these clouds may receive sunlight below the horizon and reflect them to the ground. They appear shining brightly well before dawn or after dusk.

1.8.2 Sudden Stratospheric Warming

Sudden stratospheric warming (SSW) is an event where the polar vortex of westerly winds in the northern winter hemisphere abruptly (i.e., in a few days' time) slows down or even reverses direction, accompanied by a rise of stratospheric temperature by several tens of degrees Celsius. The first observation of sudden stratospheric warming (SSW) was reported by Scherhag (1952) and the first theoretical explanation proposed by Matsuno (1971).

During northern winter, occasionally the circulation becomes highly disturbed, accompanied by a marked amplification of planetary waves. The disturbed motion is characterized by marked deceleration of zonal mean westerlies or even a reversal into zonal mean easterlies. At the same time, temperature over the polar cap

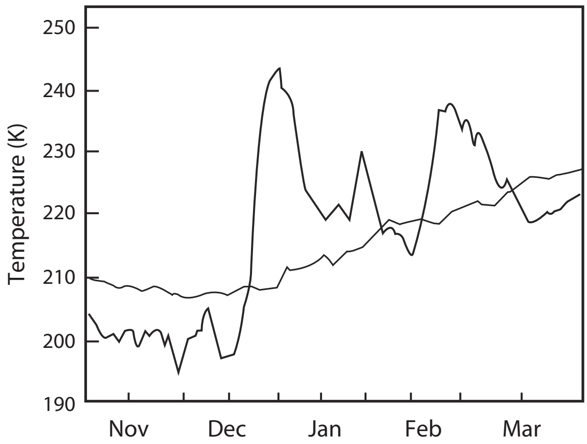

Fig. 1.22 An abrupt change in stratospheric temperature observed 90–50°N area weighted polar cap temperature at 10 hPa during a sudden stratospheric event. Bold line indicates daily changes in temperature and thin line is the mean (1958–2002) temperature from NCEP/NCAR reanalysis data sets (Adapted from A.J. Charlton and L. Polvani)

increases sharply by as much as 50 K, so the dark winter pole actually becomes warmer than the sunlit tropics. This dramatic sequence of events takes place in just a few days and is hence known as a sudden stratospheric warming.

Figure 1.22 shows the 90–50° N area averaged polar cap temperature at 10°N during November 2001 to March 2002. It can be seen that in the event of a major startospheric warming observed in December 2001, the temperature increased by about 50°C and returned back to normal after the event. In northern hemisphere winter, a few of these so-called major SSWs take place along with several minor events. Major SSWs normally occur in the northern hemisphere because orography and land–sea temperature contrasts are responsible for the generation of long (wave number 1 or 2) Rossby waves in the troposphere. These waves travel upward to the stratosphere and are dissipated there, producing the warming by decelerating the mean flow (Matsuno 1971). Since the sudden warming is observed only in the northern hemisphere, it is logical to conclude that topographically forced waves are responsible for the vertical energy propagation. The southern hemisphere with its relatively small landmasses at middle latitudes has much smaller-amplitude stationary planetary waves.

During major SSW, the north pole warms dramatically with reversal of meridional temperature gradient, and breakdown of polar vortex occurs. The polar vortex is replaced by *blocking high* over this region. The westerlies in the Arctic at 10 hPa are replaced by easterlies so that the center of the vortex moves south of 60–65°N during the breakdown of polar vortex. The vortex is either displaced entirely or split into two. This type of warming has not been observed in the Antarctic, except in 2002. Minor warming can indeed be intense and sometimes reverse the temperature gradient, but it does not result in a reversal of the circulation at 10 hPa level. This is found in the Antarctic as well as the Arctic regions.

One of the dynamic aspects of the stratospheric winter warming is the coupling effect of the warming with the tropical stratosphere. Stratospheric warming is accompanied by a slight cooling in the tropical stratosphere of both the hemispheres. This was first observed from the radiance data obtained from the Nimbus satellite (Fritz and Soules 1972). Other studies based on the satellite data also showed the presence of slight cooling in the tropical stratosphere during the warming periods (Houghton 1978). Analysis on this aspect based on rocket data from an equatorial station (Thumba, 8°N, 76°E, India) revealed occurrence of strong cooling in the equatorial stratosphere and found that the cooling penetrates to tropospheric layers especially during the peak intensity of major warming (Appu 1984). The strato-tropospheric temperatures during such occasions attain the lowest temperature of the year.

There exists a link between sudden stratospheric warmings and the QBO (Labitzke and van Loon 1999). If the QBO is in its easterly phase, the atmospheric wave-guide is modified in such a way that upward-propagating Rossby waves are focused on the polar vortex, intensifying their interaction with the mean flow. The number of warming episodes is such that it agrees reasonably well with the number of QBO cycles in the equatorial lowermost stratosphere. A possible relationship with the equatorial QBO has been noted by several authors (Holton et al. 1995; Baldwin et al. 2001). A statistically significant relationship exists between the frequeny of the occurrence of sudden stratospheric warmings in the high latitude region and the change in phases of the equatorial QBO.

1.8.3 Arctic Oscillation

Arctic oscillation (AO) is an atmospheric circulation pattern in which the atmospheric pressure over the polar regions varies out of phase with that over middle latitudes (about 45°N) on timescales ranging from weeks to decades. The oscillation extends through the entire depth of the troposphere. During late winter and early spring (January–March) it extends upward into the stratosphere where it modulates in the strength of the westerly vortex that encircles the Arctic polar cap region.

The Arctic oscillation exhibits a *negative phase* with relatively high pressure over the polar region and low pressure at midlatitudes, and a *positive phase* in which the pattern is reversed. In the positive phase, higher pressure at midlatitudes drives ocean storms farther north, and changes in the circulation pattern bring wetter weather to Alaska, Scotland, and Scandinavia, as well as drier conditions to the western United States and the Mediterranean. In the positive phase, frigid winter air does not extend as far into the middle of North America as it would during the negative phase of the oscillation. This keeps much of the United States east of the Rocky Mountains warmer than normal, but leaves Greenland and Newfoundland colder than usual. Weather patterns in the negative phase are in general opposite to those of the positive phase, as illustrated in Fig. 1.23.

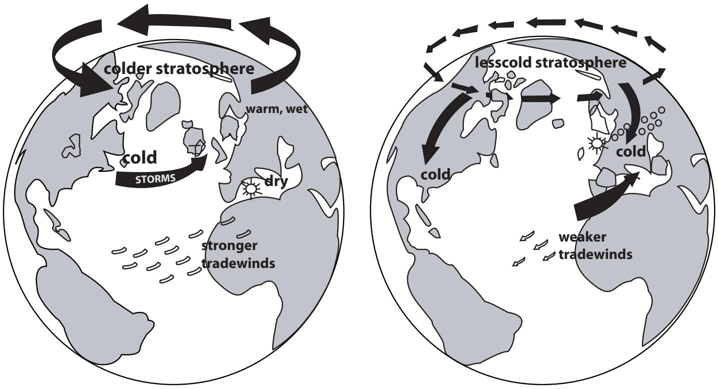

Fig. 1.23 Effects of the positive and negative phases of the Arctic Oscillation (Adapted from R. R. Stuwart 2005, Courtesy: J.M. Wallace)

In the earlier part of the last century, the Arctic oscillation alternated between its positive and negative phases. Beginning from the 1970s, however, the oscillation has tended to stay mainly in the positive phase, causing lower than normal Arctic air pressure and higher than normal temperatures in much of the United States and Northern Eurasia (Kodera and Kuroda 2000).

1.8.4 North Atlantic Oscillation

North Atlantic Oscillation (NAO) is the large-scale fluctuation in atmospheric pressure between the subtropical high pressure system located near the Azores in the Atlantic Ocean and the subpolar low pressure system near Iceland. The surface pressure drives surface winds and winter storms from west to east across the north Atlantic affecting climate from England to Western Europe as far eastward as central Siberia and the eastern Mediterranean and southward to West Africa.

In midlatitudes, NAO is the leading mode of variability over the Atlantic region. NAO is profoundly linked to the leading mode of variability of the whole northern hemisphere circulation, the Arctic oscillation. The North Atlantic oscillation and Arctic oscillation are different ways of describing the same phenomenon.

1.9 Atmospheric Tides

Atmospheric pressure, temperature, density, and winds are subject to variations with 24-hour (diurnal) and 12-hour (semidiurnal) periods. The minute but measurable variations of atmospheric parameters with lunar semidiurnal period are also caused by the gravitational attraction between the Moon and the Earth. But the variation of the atmospheric parameters with the solar diurnal and solar semidiurnal periods are caused predominantly by the heating of the atmosphere due to the absorption of solar radiation by water vapor in the troposphere and by ozone in the stratosphere and mesosphere. The heating generates pressure changes with peculiar patterns of variation with latitude, longitude, and altitude. In particular, the maximum heating rate and the associated pressure change at any given altitude travel with the subsolar point in the atmosphere, and for this reason the tides generated by solar heating are known as *migrating tides*.

The migrating meridional and zonal pressure gradients generate accelerations of air parcels which are subject to Coriolis forcing as they progress. In addition to the hydrostatic balance, mass continuity, and thermodynamic energy conservation, the ensuing equilibrium between global distribution of pressure and velocity fields associated with the tides is subject to the principal boundary conditions of the Earth's poles. The natural boundaries provided by the poles result in specific modes of oscillations with specific latitudinal structures only being possible for each tidal period.

Basically tides are generated by the daily variation of the atmospheric heating by solar radiation. The heating is generated by the absorption of solar UV radiation by ozone in the stratosphere and mesosphere, and the absorption of near-infrared bands by water vapor in the troposphere.

1.10 Major Greenhouse Gases in the Troposphere and Stratosphere

Atmospheric ozone, water vapor, and carbon dioxide are the major greenhouse gases which are abundant in the atmosphere and control the radiation balance of the Earth atmosphere system. These greenhouse gases are very good absorbers in the infrared part of the spectrum and regulate the temperature of the atmosphere. In this section the characteristics and distribution of these major greenhouse gases are discussed.

1.10.1 Stratospheric Ozone

Stratospheric ozone is the most important minor constituent present in the Earth's atmosphere. The more or less continuous increase in temperature with height in the stratosphere is mainly due to the absorption of solar ultraviolet radiation by a layer of ozone molecules with peak abundance near 25 km. Although ozone is a minor constituent in the atmosphere, it absorbs ultraviolet radiation very effectively at wavelengths between 200 nm and 300 nm. This property of the ozone protects the life on Earth by preventing the harmful radiation reaching the Earth's surface.

The word *ozone* is derived from the Greek word *ozein*, meaning to smell. Ozone has a pungent odour that allows ozone to be detected even in very low amounts. Ozone will rapidly react with many chemical compounds and is explosive in concentrated amounts.

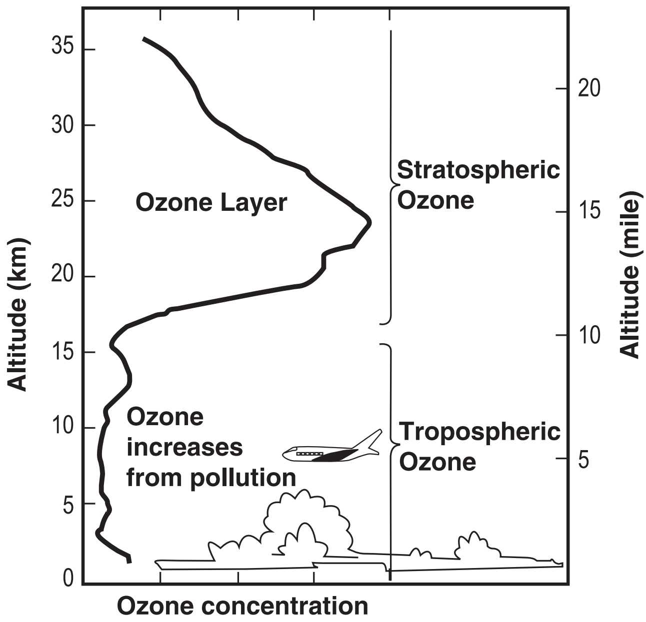

Fig. 1.24 Vertical distribution of Atmospheric ozone (Adapted from : WMO 2007)

Vertical distribution of ozone in the atmosphere is shown in Fig. 1.24. More than 90% of ozone resides in the stratosphere in what is commonly known as the *ozone layer* or *ozonosphere*. The remaining ozone is found in the troposphere.

Ozone molecules have a relatively low abundance in the atmosphere. In the stratosphere near the peak of the ozone layer, there are up to 12,000 ozone molecules for every billion air molecules. Most air molecules are in the form of molecular oxygen (O_2) or molecular nitrogen (N_2). Near to the Earth's surface, ozone is even less abundant, with a typical range of 20–100 ozone molecules formed in each billion air molecules. The highest surface ozone values are due to the formation of ozone in polluted air by anthropogenic activities.

The absorption of UV_b radiation by ozone is a source of heat in the stratosphere. This helps to maintain the stratosphere as a stable region of the atmosphere with temperatures increasing with altitude. As a result, ozone plays a key role in controlling the temperature structure of Earth's atmosphere.

Stratospheric ozone is considered good for humans and other lifeforms because it absorbs UV_b radiation from the Sun. Otherwise, UV_b would reach Earth's surface in amounts that are harmful to a variety of lifeforms. In humans, as their exposure to UV_b increases, so does their risk of skin cancer, cataracts, and a suppressed immune system. Excessive UV_b exposure also can damage terrestrial plant life, single-cell organisms, and aquatic ecosystems. Other UV radiation, UV_a, which is not absorbed significantly by ozone, causes premature aging of the skin (WMO 2007).

Ozone is also formed near Earth's surface in natural chemical reactions and in reactions caused by the presence of man-made pollutant gases. Ozone produced by pollutants is *bad* because more ozone comes in direct contact with humans, plants, and animals. Increased levels of ozone are generally harmful to living systems because ozone reacts strongly to destroy or alter many other molecules. Excessive ozone exposure reduces crop yields and forest growth. In humans, ozone exposure can reduce lung capacity, and cause chest pains, throat irritation, and coughing, thereby worsening preexisting health conditions related to the heart and lungs. In addition, increases in tropospheric ozone lead to a warming of Earth's surface. The negative effects of increasing tropospheric ozone contrast sharply with the positive effects of stratospheric ozone as an absorber of harmful UV_b radiation from the Sun.

Ozone is a natural component of the clean atmosphere. In the absence of human activities on Earth's surface, ozone would still be present near the surface and throughout the troposphere and stratosphere. Ozone's chemical role in the atmosphere includes helping to remove other gases, both those occurring naturally and those emitted by human activities. If all the ozone were to be removed from the lower atmosphere, other gases such as methane, carbon monoxide, and nitrogen oxide would increase in abundance.

Ozone is constantly being produced and destroyed in a natural cycle. However, the overall amount of ozone is essentially stable. Large increases in stratospheric chlorine and bromine have upset that balance. Therefore, ozone levels are beginning to fall. The ozone depletion process begins when chloroflurocarbons (CFCs) and other ozone-depleting substances (ODS) leak or are released from equipments. Winds efficiently mix the troposphere and evenly distribute the gases. CFCs are extremely stable, and they do not dissolve in rain. After a period of a few years, ODS molecules reach the stratosphere. Strong UV light breaks apart the ODS molecule. CFCs release chlorine atoms and halons release bromine atoms. It is these atoms that actually destroy ozone, not the intact ODS molecule. It is estimated that one chlorine atom can destroy over 100,000 ozone molecules before finally being removed from the stratosphere.

Ozone photochemistry, transportation, ozone hole, and their influence on tropospheric weather systems are discussed in detail in later chapters of this book.

1.10.2 Carbon Dioxide

Carbon dioxide has a relatively constant mixing ratio with height in the atmosphere, and is more or less evenly distributed. The main sources of carbon dioxide are burning of fossil fuels, human and animal respiration, the oceans, and volcanic activity. The main sinks are photosynthesis and the production of carbonates (limestones) in the ocean/land system. The rate of removal of carbon dioxide is observed to be less than the generation (from fossil fuel burning) because the concentration of carbon dioxide in the atmosphere has been rising steadily during the 20th century as illustrated in Fig. 1.25.

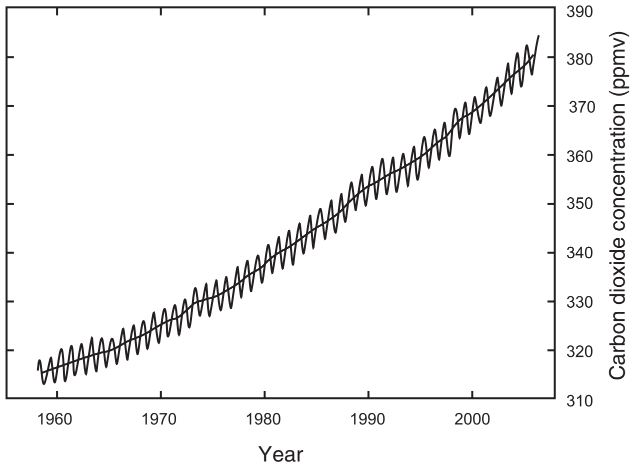

Fig. 1.25 Carbon dioxide concentration in the Earth's atmosphere, observed at Mauna Loa, Hawaii (Courtesy: Robert A. Rhode, Global Warming Art)

About 99% of the Earth's carbon dioxide is dissolved in the oceans. Because solubility is temperature-dependent the gas therefore enters or leaves the oceans. It is estimated that the annual amount of carbon dioxide entering or leaving the air by all mechanisms is about one tenth of the total carbon dioxide content of the atmosphere.

1.10.3 Water Vapor

Water vapor is unique among atmospheric trace constituents in that conditions for saturation are common in the atmosphere. This property is the most important factor governing the distribution of water vapor in the atmosphere, both in the troposphere, where it varies by as much as four orders of magnitude in a vertical profile, and in the stratosphere, where variations are much smaller but still significant (see Fig. 1.26).

Water vapor is extremely important in radiative absorption and emission processes in the atmosphere. Its concentration is highly variable. Although always present, in some localities it is difficult to measure, but in the tropics its concentration can be as high as 3% or 4% by volume. Water vapor content of air is a strong function of air temperature. For example, air at 40°C can hold up to 49.8 g of water per kilogram of dry air, while at 5°C this reduces to 5.5 g per kilogram of dry air.

The release of latent heat from condensation of water in the atmosphere is significant in the global energy budget and climate. Relatively small amounts of water vapor can produce great variations in weather. This is largely due to changes in its concentration and in latent heat release, particularly below 6 km where a high proportion of moisture lies.

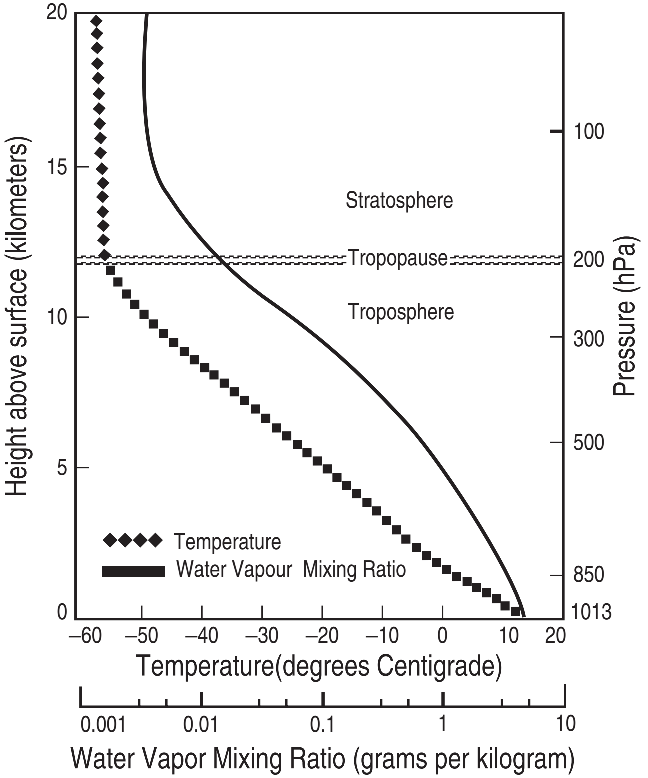

Fig. 1.26 Vertical distribution of water vapor (Courtesy: D. Siedel)

The major sources of water vapor are evaporation from water surfaces and transpiration from plant life. The main sink is condensation in clouds with resulting precipitation over oceans and land. On an average the concentration of atmospheric water vapor decreases with altitude, although this distribution may vary with space and time.

1.10.4 Water Vapor in the Stratosphere

In the stratosphere above 100 hPa, the distribution of water vapor can be explained as a balance between dry air entering through the tropical tropopause and a source of water vapor from methane oxidation in the upper stratosphere. The stratospheric circulation helps to determine the distribution, along with wave-induced mixing and upward extension of tropospheric circulation. Variations in the zonal direction are rapidly mixed so that water vapor is nearly constant following a fluid element. Nearly all air passing from the troposphere to the part of the stratosphere above

100 hPa enters through the tropical tropopause, where the removal of water vapor by low temperatures and a combination of other processes dries the air to around 3.5–4 ppmv in the annual mean. Some of this dry air rises slowly in the tropics, but most spreads poleward, or is mixed with midlatitude air, especially in the lowest few kilometers of the stratosphere. Consequently, water vapor concentrations increase upward and away from the equator as methane is oxidized.

Below 100 hPa, the extratropical lower stratosphere is moistened by seepage from the troposphere, mostly by roughly horizontal transport across the subtropical tropopause. This horizontal transport is stronger in the summer hemisphere. There is also a hemispheric asymmetry in the transport, with the south Asian monsoon in boreal summer significantly moistening the northern hemisphere, more than similar monsoon circulation in the southern hemisphere. Other important seasonal variations in the stratosphere occur in the winter and spring polar vortex, especially in the Antarctic, where cold temperatures cause dehydration via the formation of ice clouds, which play a crucial role in catalytic ozone destruction in every spring.

In the tropics and subtropics, upper-tropospheric water vapor is strongly influenced by the meridional Hadley circulation and the zonal Walker circulation. The predominant source for moisture in the tropical and subtropical upper troposphere is convection. In general, moist areas appear in the convective areas over the western Pacific, South America, and Africa. Moist areas also appear seasonally in the region of the Asian summer monsoon and along the intertropical and South Pacific convergence zones.

The seasonality of surface temperature and of convection, which roughly follow the Sun, as well as seasonal variations in monsoon circulation produce associated seasonal changes in water vapor in the troposphere. This relationship between convection and upper tropospheric moisture changes sign near the tropical tropopause, somewhere between 150 hPa and 100 hPa, so that convection dries the tropopause region. Water vapor is also influenced by fluctuations at both shorter and longer timescales, including the quasi-biennial oscillation in the stratosphere, and El Niño and the southern oscillation and the tropical intraseasonal oscillation in the troposphere.

Water vapor is highly variable in middle to high latitudes in the upper troposphere, and can be supplied by transport from the tropics, by mesoscale convection, or by extratropical cyclones. Dry air can be transported from the subtropics or from the extratropical lower stratosphere. These transport phenomena tend to be periodic rather than continual.

A complex mix of processes takes place at the tropical tropopause to remove water vapor from air, as it enters the stratosphere. Within the background of large-scale mean ascent, the dehydration processes include smaller-scale ascent, radiative and microphysical processes within clouds, and wave-driven fluctuations in temperature. The location, strength, and relative importance of these processes vary seasonally. But the seasonal variation in tropopause level water vapor is influenced by the seasonal variation in tropical tropopause temperatures. Air rising through the tropopause is marked with seasonally varying mixing ratio, and retains these markings as it spreads rapidly poleward and more slowly upward into the stratosphere.

1.11 Upper Troposphere and Lower Stratosphere

The upper troposphere and lower stratosphere (UT/LS) is a complex region and the exchange process take places through the interface between the troposphere and stratosphere. Since the radiative and chemical timescales are relatively long, the transport is highly significant. The radiative properties and phase changes of atmospheric moisture link the hydrological and energy cycles of the Earth system. As the average residence time of water vapor in the atmosphere is around 10 days, the atmospheric branch is a relatively fast component of the global hydrological cycle. Tropical rainfall is particularly important as a forcing mechanism for the atmospheric large-scale circulation and climate.

The UT/LS has some distinct characteristics that influence Earth's climate (see Fig. 1.27). The pressures are still high enough to influence the course of reactions and photochemical processes. This region is below the ozone layer such that the radiation is limited to wavelengths greater than ~290 nm. Upper troposphere and lower stratosphere contain the coldest parts of the lower atmosphere, to the extent that highly reactive particles can be produced. These particles facilitate heterogeneous reactions taking place on a solid substrate and multiphase reactions happening in a liquid droplet, thereby altering the composition of this region. The particles,

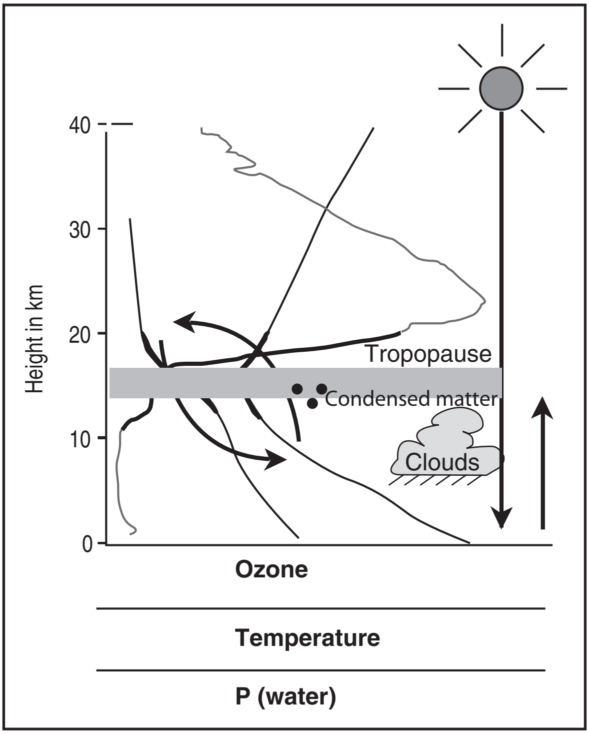

Fig. 1.27 Issues of UT/LS region (Adapted from A.R. Ravishankara and T. Shepherd 2001)

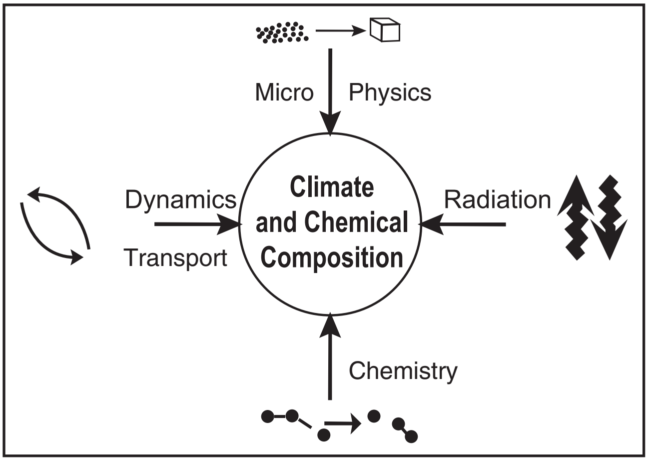

Fig. 1.28 UT/LS Region-complex interplay between dynamics, transport, radiation, chemistry, and microphysics (Adapted from A.R. Ravishankara and T. Shepherd 2001)

especially cirrus clouds, also directly interact with radiation. However, these are the regions where a clear separation of the timescales for chemistry and dynamical transport does not exist. Thus, chemical, microphysical, and dynamical processes all play an important role in the determination of ozone abundance and the radiative balance of the atmosphere.

Thus UT/LS region involves a complex interplay between dynamics, transport, radiation, chemistry, and microphysics. Dynamics and radiation lead to the low temperatures that form condensed matter through microphysical processes as represented in Fig. 1.28. Microphysics in turn affects chemistry, as do temperatures, solar radiation, and transport of chemical species, and chemistry sequentially feeds back onto climate through radiation (Ravishankara and Liu 2003).

Many climate and environmental quality issues are involved in the interface between the stratosphere and the troposphere. The UT and LS are inseparably connected via transport of chemicals and their mutual interactions are very large and significant. The source of all the ingredients for photochemical production of ozone in the upper troposphere and destruction in the lower stratosphere originates from the lower atmosphere and has to pass through the upper troposphere to reach the lower stratosphere. Similarly, the contents of the lower stratosphere are passed through the upper troposphere to be removed, and thereby affect this sensitive area (Ravishankara and Liu 2003).

Models developed in the UT/LS region have not yet performed well. Our present-day understanding of trends in radiatively important trace species, such as ozone and water vapor, is not adequate. Information regarding the role of the stratosphere in climate requires the proper treatment of transport and mixing in the upper troposphere and lower stratosphere, radiative processes, and the chemistry and microphysics of stratospheric ozone depletion, all of which are coupled since their timescales are similar. Moreover, atmospheric measurements in this region are somewhat difficult and incomplete.

1.12 Atmospheric Aerosols

Atmospheric aerosols are small airborne particles of widely differing chemical composition. They can either be of natural or of anthropogenic origin and it is estimated that anthropogenic aerosols constitute around 50% of the global mean aerosol optical thickness. The major sources of anthropogenic aerosols are from fossil fuel burning and biomass burning.

Atmospheric aerosols are important in many ways. Aerosol content affects the Earth's albedo and plays a major role in the global radiation balance and climate. Various aerosols act as cloud condensation nuclei and are important in the formation of clouds and precipitation. In addition, the scattering property of the aerosols can be used by a number of next-generation active remote-sensing instruments in derivation of geophysical parameters.

Most aerosol particles originate from blowing soil, smoke, volcanoes, and the oceans. Particles made of sodium chloride or magnesium chloride are hygroscopic and therefore act as good sites for the condensation of water to form cloud droplets. The concentration of the aerosols varies considerably but is typically on the order of 103 cm^{-3} over oceans, 104 cm^{-3} over rural land, and 105 cm^{-3} over cities. The concentrations generally decrease with altitude.

The size of aerosol particles is usually given as the diameter of the particle assuming a spherical shape. The sizes of different aerosol particles in the atmosphere are illustrated in Fig. 1.29. Aerosols are usually assigned into three size categories: (i) aitken particles, or *nucleation mode* (0.001–0.1 μm diameter); (ii) large particles, or *accumulation mode* (0.1–1 μm diameter); and (iii) giant particles, or *coarse particle mode* (>1 μm diameter).

The terms *nucleation mode* and *accumulation mode* refer to the mechanical and chemical processes by which aerosol particles in those size ranges are usually produced. The smallest aerosols, in the nucleation mode, are principally produced by gas-to-particle conversion, which occurs in the atmosphere. Aerosols in

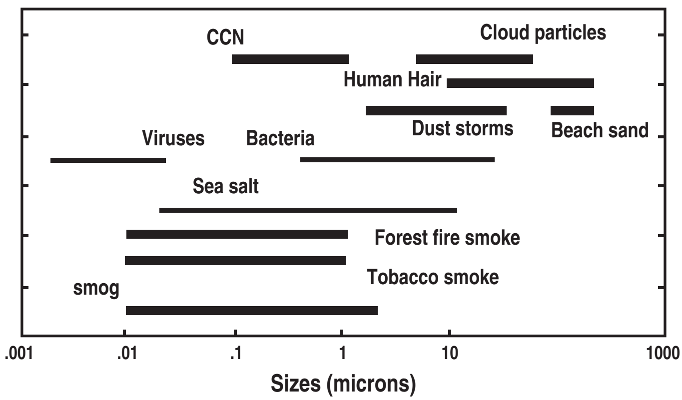

Fig. 1.29 Size of different aerosols (Courtesy: Bruce Caron, New Media Studio)

the accumulation mode are generally produced by the *coagulation* of smaller particles and by the *heterogeneous condensation* of gas vapor onto existing aerosol particles.

1.12.1 Water-soluble Aerosols

Aerosols consist of water-soluble compounds, such as sulphate, nitrate, and sea salt, and are efficient cloud condensation nuclei (CCN). In unpolluted continental conditions, smaller particles are more likely to be water-soluble. Nearly 80% of the particles in the 0.1–0.3 μm size range are comprised of water-soluble particles. Over oceans, however, much of the coarse particle mode comprised of sea salt aerosols is water-soluble. Water-soluble aerosols are *hygroscopic* and they are capable of attracting water vapor from the air. The size of hygroscopic particles varies with relative humidity, leading to changes in optical properties as well. The presence of polar functional groups on organic aerosols, particularly carboxylic and dicarboxylic acids, makes many of the organic compounds in aerosols water-soluble and allows them to participate in cloud droplet nucleation. Aerosols, such as metal oxides, silicates, and clay minerals, originate from soil dust or volcanoes and are insoluble.

1.12.2 Residence Time of Aerosols

The residence time of aerosols depends on their size, chemistry, and height in the atmosphere. Particle residence times range from minutes to hundreds of days. Aerosols between 0.1 and 1.0 μm, known as the accumulation mode, remain in the atmosphere for a longer time. Smaller aerosols (the nucleation mode) are subject to Brownian motion. As a result, higher rates of particle collision and coagulation increase the size of individual particles and remove them from the nucleation mode. The coarser particles (>1 μm radius) have higher sedimentation rates and therefore the residence time is lower.

Aerosol lifetime is on the order of a few days or weeks, and aerosols are produced unevenly on the surface of the Earth. Variation between different parts of the globe in optical thickness and radiative forcing due to aerosols, can amount to tens of W m^{-2}, which is a magnitude higher than the global means for these parameters.

1.12.3 Tropospheric Aerosols

Tropospheric aerosols are removed fairly rapidly by deposition or rainout and typically have residence times of about 1 week. Their spatial distribution is therefore very inhomogeneous and strongly correlated with the source regions. Present knowledge on the optical properties of tropospheric aerosols, as well as their spatial and temporal evolution, is inadequate. Routine satellite monitoring of tropospheric aerosols, when not obstructed by clouds, should provide insights into not only these aspects but also the way aerosols alter cloud optical properties.

1.12.4 Stratospheric Aerosols

Stratospheric aerosols have much longer residence times, on the order of about 1 year, and therefore have a more uniform distribution. As a result, volcanic explosions in which the debris reaches into the stratosphere can perturb global climate for several years. Satellite and sonde measurements indicate a steady increase in the background concentration of stratospheric sulphate of about 40–60% over the last decade. It seems that this record is not affected by volcanic aerosols but could reflect an influx of increasing tropospheric aerosols. Emissions from subsonic aircraft flying in the stratosphere are also contributing aerosols in the stratosphere. Considering the expected growth in future air traffic, it is indeed necessary to regularly monitor the stratospheric aerosol trends on a global basis.

Problems and Questions

1.1. Find the global mean surface pressure in hPa, if the mass of the atmosphere is 5.10×10^{18} kg. Assume that the mean value of the acceleration due to gravity is 9.807 m s^{-2}. How much is the global mean surface pressure different from the standard mean sea level pressure?

1.2. Determine the mean molecular weight of a sample of air consisting of nitrogen, oxygen, and argon, given that the molecular weight of nitrogen, oxygen, and argon is 28.01, 32.00, and 39.85, respectively.

1.3. An inflated balloon at the ground is taken in an aircraft which is flying at 8 km height. If the cabin is not pressurized, what would happen to the balloon? Explain.

1.4. What is the pressure at the top of the Eiffel Tower roof, if the pressure and density at the surface are 1,013.25 hPa and 1.225 kg m^{-3}, respectively. (Hint: Eiffel Tower roof height is 330 m)

1.5. Estimate the height at which the density falls to 0.5 kg m^{-3}, and the pressure decreases to 5 hPa. Assume that the pressure and density decrease exponentially with a scale height of 7.5 km, given that surface level pressure is 1,013 hPa and density is 1.225 kg m^{-3}.

1.6. Let the mean lapse rate of the troposphere be 6.5°C km^{-1}. If the surface temperature is 30°C, what is the temperature at the top of Mt. Everest, which is 8,848 m above the surface?

1.7. A hot air balloon is traveling eastward along 30°N at a mean speed of 12 m s^{-1} in half part of the globe, and 16 m s^{-1} in the remaining portion. If the prevailing wind is calm, estimate how much time it takes to circumnavigate the entire globe at this latitude belt.

1.8. Consider an aircraft flying at a height of 12 km from Paris to Tokyo at a speed of 800 km per hour. If the aircraft enters into the centre of the subtropical jet stream, which has an average core speed of 50 m s^{-1}, estimate the time gained by the aircraft on reaching the destination. The distance between Paris and Tokyo is approximately 10,000 km.

1.9. The temperature of the equatorial tropopause at 18 km is −87°C and that at 30°N is −66°C at 14 km altitude. Calculate the temperature gradient between the equator and 30°N. Neglect the curvature of the Earth and assume mean lapse rate of 6.5°C.

1.10. In the equatorial region, on January 2007, the zonal wind at 30 km is westerly at a speed of 15 m s^{-1}. In January 2008, what would be phase of the zonal wind at 30 km and at 18 km? Explain.

1.11. It is assumed that during the early stages of its formation the Earth's atmosphere contained a large amount of hydrogen, but the present atmosphere does not have very much hydrogen gas in it. Where did that hydrogen go?

1.12. The tropopause temperature in the tropics is much colder than the tropopause temperature in middle and polar latitudes. What structural aspect of the atmospheric temperature profile makes this possible even though the surface temperature in the tropics is much warmer than the surface temperature in middle and polar latitudes?

1.13. What are the general latitudinal trends in temperatures as you move poleward from the equator in troposphere and stratosphere? Explain whether the tropospheric changes are more severe in the summer or winter hemisphere?

1.14. Why is the height of the tropopause different at different latitudinal zones? Suppose the height of the tropopause is same throughout the globe, what would be the wind structure and temperature pattern of the lower atmosphere?

1.15. If the tropical tropopause is raised to 25 km, discuss the changes that would occur in the tropospheric weather systems. Describe the role of tropopause in maintaining the general circulation of the atmosphere.

1.16. What would be the vertical thermal structure of the atmosphere in the absence of ozone? Why is stratospheric ozone called *good ozone* and tropospheric ozone called *bad ozone*?

1.17. In the Earth's atmosphere, the maximum ozone concentration is noted around 25 km altitude region. Why then is stratopause found at 50 km, and not at 25 km?

1.18. How does the stratospheric northern hemisphere winter temperature structure compare to that of the southern hemisphere winter? What are regions of coldest temperatures in the stratosphere termed?

1.19. What latitude regions experience least and greatest changes in zonal winds? Explain. What is the main driving mechanism responsible for the Earth's large-scale atmospheric circulations?

1.20. What causes semiannual oscillations in the tropical atmosphere? Why are they absent at higher latitudes?

1.21. Describe the charactertics of stratospheric warming occurrence in the northern and southern hemispheres. Does this high latitude phenomenon have influence on other altitude as well as latitude regions? Explain.

1.22. What are the various sources of aerosols in the atmosphere? How do aerosols reach the stratosphere? Suppose you remove the entire aerosols from the atmosphere, what would happen?

References

Andrews DG, Holton JR, Leovy CB (1987) Middle Atmosphere Dynamics, Academic, New York

Asnani GC (2005) Tropical Meteorology, Second Edition, Vol. 2, Pune, India

Appu KS (1984) On perturbations in the thermal structure of tropical stratosphere and mesosphere in winter, Indian J Radio Space Phys, 13: 35–41

Baldwin MP, Gray LJ, Dunkerton TJ, Hamilton K, Haynes PH, Randel WJ, Holton JR, Alexander MJ, Hirota I, Horinouchi T, Jones DBA, Kinnersley JS, Marquardt C, Sato K, Takahashi M (2001) The quasibiennial oscillation, Rev of Geophys, 39: 179–229

Baldwin MP, Stephenson DB, Thompson DWJ, Dunkerton J, Charlton AJ, ONeil A (2003) Stratospheric memory and skill of extended range weather forecasts, Science, 301: 636–640

Bethan S, Vaughan G, Reid SJ (1996) A comparison of ozone and thermal tropopause heights and the impact of tropopause definition on quantifying the ozone content of the troposphere, Quart J Royal Met Soc, 122: 929–944

Brasseur G, Solomon S (1984) Aeronomy of the Middle Atmosphere, D Riedel Publishing Company, Dordrecht

Caron B, What are aerosols, New Media Studio, The National Science Digital Library (*http://www.newmediastudio.org/DataDiscovery/Aero_Ed_Center/Charact/size_of_aerosols.gif*)

Charlton A, Polvani L (2006) Algorithm for identifying sudden stratospheric warming, Dept. of Applied Physics and Applied Mathematics, Columbia University

EUMeTrain, Vertical cross sections, PV anomaly (*http://www.zamg.ac.at/eumetrain/CAL_Modules/VCS/Content/images/pv2.jpg*)

Frits S, Soules S (1972) Planetary variations of stratospheric temperatures, Mon Wea Rev, 100: 582–589

Geerts B, Linacre E (1997) The height of the tropopause, University of Wyoming (*http://www-das/~geerts/cwx/notes/chap01/trop-height01.gif*)

Gettelman A, Forster PM. de F, (2002) A climatology of the tropical tropopause layer, J Met Soc Japan, 80 (4B), 911–924

Hamilton K, Hertzog A, Vial F, Stenchikov G (2004) Longitudinal variation of the stratospheric quasi-biennial oscillation, J Atmos Sci, 61: 383–402

Hartjenstein G (1999) Tropopause folds and the related stratosphere troposphere mass exchange, Project Report of University of Munich (*http://www.lrz-muenchen.de/projekte/hlr-projects/1997-1999/cd/daten/pdf/uh22102.pdf*)

Haynes P, Shepherd T (2001) Report on the SPARC Tropopause Workshop, Bad Tolz, Germany, 17–21 April 2001

Hays J, deMenocal P, Lecture Notes on the Seasonality of ozone, Section 1.5, Columbia University (*http://www.ideo.columbia.edu/edu/dees/V1003/lectures/ozone/index.html*)

Hirota I, Yasuko H (2000) Interannual variations of planetary waves in the Southern Hemisphere stratosphere, 2nd SPARC General Assembly 2000, 26–30 October, Mardel Plata, Argentina

Holton JR, Haynes PH, McIntyre ME, Douglass AR, Rood RB, Pfister L (1995) Stratosphere-troposphere exchange, Rev Geophys, 33: 403–440

Houghton JT (1978) The stratosphere and mesosphere, Quart J Roy Met Soc, 104: 1–29

Kodera K, Kuroda Y (2000) Tropospheric and stratospheric aspects of the Artctic Oscillation, Geophys Res Lett, 27: 3349–3352

Labitzke K, van Loon H (1999) The Stratosphere, Phenomena, History and Relevance, Springer, Berlin

Marquardt C, Naujokat B (1997) An update of the equatorial QBO and its variability. 1st SPARC Gen. Assembly, Melbourne Australia, WMO/TD-No. 814: 87–90

Matsuno T (1971) A dynamical model of stratospheric warmings, J Atmos Sci, 28: 1479–1494

McPhaden MJ, Busalacchi AJ, Cheney R, Donguy J, Gage KS, Halpern D, Julian M Ji, Meyers PG, Mitchum GT, Niiler PP, Picaut J, Reynolds RW, Smith N, Takeuchi K (1998) The tropical ocean-global atmosphere observing system: A decade of progress, J Geophys Res, 103: 14169–14240

National Weather Service, Northern hemisphere cross section showing jetstreams and tropopause elevations, NOAA (*http://www.srh.noaa.gov/jetstream/global/image/jetstream3.jpg*

Randel WJ, Udelhofen P, Fleming E, Geller MA, Gelman M, Hamilton K, Karoly D, Ortland D, Pawson S, Swinbank R, Wu F, Baldwin M, Chanin ML, Keckhut P, Labitzke K, Remsberg E, Simmons A, Wu D (2004) The SPARC Intercomparison of Middle Atmosphere Climatologies. J Climate, 17: 986–1003

Ravishankara AR, Liu S (2003) Highlights from the Joint SPARC-IGAC Workshop on Climate-Chemistry Interactions, held at Giens, France, during April 02–06, 2003

Ravishankara AR, Shepherd T (2001) Upper tropospheric and lower stratospheric processes, SPARC/IGAC Activities, SPARC Brochure – The SPARC Initiative

RMIT University, Jetstreams, Current Global Climates (*http://users.gs.mit.edu.caa/global/graphics/jetstreams.jpg*)

RMIT University, Vertical circulation cells, Current Global Climates (*http://users.gs.rmit.edu.au/caa/global/graphics/hadley.jpg*)

Russel CT (1987) The magnetosphere, in The Solar Wind and the Earth, edited by S.-I Akasofu and Y. Kamide, Terra Scientific Publishing Co., Tokyo

Shapiro MA, Hampel T, Kruger AJ (1987) The tropopause fold, Mon Wea Rev, 115: 444–454

Sherhag R (1952) Die explosion sartigen stratospharener warmungen des spatwinters 1951–1952, Ber Deut Wetterd, 6: 51–53

Siedel D, Annual cycles of tropospheric water vapor, Air Resource Laboratory, NOAA

Stuwart RR (2005) The oceanic influence on North American drought, in Oceanography in the 21st Century – An Online Textbook, Department of Oceanography, Texas A&M University

Tameeria (2007) Oxygen buid up in Earth's atmosphere, Wikipedia (*http://en.wikipedia.org/wiki/User.tameeria/images*)

The Green Lane, Environmental Canada (*http://www.qc.ec.gc.ca/Meteo/images/Fig_5-9_a.jpg*)

University of Leicester, Medium frequency radio frequency theory notes (*http://www.k1ttt.net/technote/kn41f/kn41f8_files/ionosphereprofile.gif*)

US Standard Atmosphere (1976) US Government Printing Office, Washington DC

Wallace JM, Hobbs PV (2006) Atmospheric Science – An Introductory Survey, Second edition, Elsevier, New York

Wirth V (2003) Static stability in the extratropical tropopause region. J Atmos Sci, 60:1395–1409

Wirth V, Szabo T (2007) Sharpness of the extratropical tropopause in baroclinic life cycle experiments, Geophys Res Letts, 34: L02809, doi:10.1029/2006GL028369

WMO (2007) Scientific Assessment of Ozone Depletion : 2006, Global Ozone Research and Monitoring Project, WMO Report No. 50, Geneva

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