RADIATION PROCESSES IN THE LOWER AND MIDDLE ATMOSPHERE

By

Chalchisa Daba Iticha

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Advisor: ______________________
Dr.Yitagessu Elfaged

Examiners: ______________________
Dr. Lemi Demeyu

Examiners: ______________________
Dr. Feraol Fana

ADDIS ABABA UNIVERSITY

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Author: **Chalchisa Daba Iticha**

Title: **RADIATION PROCESSES IN THE LOWER AND MIDDLE ATMOSPHERE**

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Abstract

The movement and causes of this movement in Earths atmosphere depend merely on the magnitude and distribution of the net radiative heating of the earth atmosphere system. In the troposphere, the net adiabatic heating rate is dominated by the imbalance between the transfer of heat from the surface and the thermal emission of radiation to space. Latent heat is the major component of the flux from the surface to the atmosphere, and clouds play a major role in the emission of radiation to space. In the stratosphere the net heating depends solely on the imbalance between local absorption of solar UV radiation and infrared radiative loss.
In this region, ozone is the principal absorber and carbon dioxide is the dominant emitter. Infrared emission by ozone and water vapor, molecular oxygen, carbon dioxide and nitrogen dioxide play a secondary roles. The distribution of the radiative sources and sinks due to the above gases exerts a zero order control on the large-scale seasonally varying mean temperature and zonal wind fields of the stratosphere.

These radiative process are therefore, significant to understand the stratosphere- troposphere interactions. In generally, this study (project) introduces the basic principles of radiation, solar radiation and their process in the stratosphere and troposphere.

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Chapter 1

Introduction

Radiation is the process through which energy moves through space from the source, without any material medium. Radiation sources are generally collections of matter or devices that convert other forms of energy into radiative energy [8]. In some cases the energy to be converted is stored within the object like the Sun and radioactive materials. In other cases the radiation source is only an energy converter, and other forms of energy must be applied in order to produce radiation. Most forms of radiation can penetrate through a certain amount of matter. But in most situations, radiation energy is eventually absorbed by the material and converted into another form of energy.

Solar radiation is composed of electromagnetic waves that travel through space. Electromagnetic waves are formed when an electric field couples with a magnetic field. The electric and magnetic fields of an electromagnetic wave are perpendicular to each other and to the direction of wave. Electromagnetic energy is a form of energy that is transferred by radiation from all things in nature. Electromagnetic energy spectrum comprises a broad band of wavelengths that travels from its source through space in the form of harmonic waves at the uniform speed of light in vacuum ($3 \times 10^8 ms^{-1}$). Radiation is the term that relates to the emission and propagation of electromagnetic energy in the form of waves [9].

In other words, radiation may be defined as a process in which energy is transmitted across space whether or not a material medium (as for conduction, convection, or advection) is present.

Approximate wavelengths, frequencies and energy levels of the various regions of the electromagnetic spectrum are listed in Table 1.1. The various types of electromagnetic radiation in order of increasing energy level are also illustrated in Table 1.1. Radio waves are the region of the electromagnetic spectrum with very long wavelengths ($> 0.1m$). Radio, TV, and radar communications occur using these types of waves. Microwaves are the spectrum lying between ultrahigh frequency (UHF) radio waves and (heat) infrared waves. Microwaves are used to generate heat and for communications. They may cause heat damage to tissues .

	Wavelength (m)	Frequency (Hz)	Energy (J)
Radio	$> 1 \times 10^{-1}$	$< 3 \times 10^{9}$	$< 2 \times 10^{-24}$
Microwave	$1 \times 10^{-3} - 1 \times 10^{-1}$	$3 \times 10^{9} - 3 \times 10^{11}$	$2 \times 10^{-24} - 2 \times 10^{-22}$
Infrared	$7 \times 10^{-7} - 1 \times 10^{-3}$	$3 \times 10^{11} - 4 \times 10^{14}$	$2 \times 10^{-22} - 3 \times 10^{-19}$
Optical	$4 \times 10^{-7} - 7 \times 10^{-7}$	$4 \times 10^{14} - 7.5 \times 10^{14}$	$3 \times 10^{-19} - 5 \times 10^{-19}$
UV	$1 \times 10^{-8} - 4 \times 10^{-7}$	$7.5 \times 10^{14} - 3 \times 10^{16}$	$5 \times 10^{-19} - 2 \times 10^{-17}$
X-ray	$1 \times 10^{-11} - 1 \times 10^{-8}$	$3 \times 10^{16} - 3 \times 10^{19}$	$2 \times 10^{-17} - 2 \times 10^{-14}$
Gamma-ray	$< 1 \times 10^{-11}$	$> 3 \times 10^{19}$	$> 2 \times 10^{-14}$

URL: http://imagine.gsfc.nasa.govfc.nasa.gov/docs/science/know_l1/spectrum_chart.html

Table 1.1: Wavelength, frequency, and energy of various regions in the electromagnetic spectrum [10]

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Chapter 2

Energy Of Radiation

Electromagnetic waves consist of radiant energy. The quantity of radiant energy may be calculated by integrating radiant flux with respect to time. It is generally thought of as radiation emitted by a source into the surrounding environment. Specific forms of radiant energy include electron space discharge, visible light, vacuum energy, and other wave types. Radiant energy is exhibited in the spontaneous nuclear disintegration with emission of particulate or electromagnetic radiations [8, 10]. A photon is an elementary particle which is the carrier of electromagnetic radiation of all wavelengths. Its most important characteristic is the quantity of energy it contains. The photon energy is the energy of the individual photons that determines the type of electromagnetic radiation, such as light, X-ray, radio signals, etc. One of the significant characteristics of photon energy is that it generally determines the penetrating ability of the radiation. The lower-energy X-ray photons are often referred to as soft radiation, whereas those at the higher-energy end of the spectrum would be termed as hard radiation. Generally, high-energy (hard) X-ray photons are more penetrating than the softer portion of the spectrum [10].
The energy associated with a photon is given by

$$E = hv = hc/\lambda, \tag{2.1}$$

Where h is the Planck's constant ($6.626 \times 10^{-26} Js^{-1}$), ν is the frequency of light, λ is the wavelength, c is speed of light. So the energy of a photon varies inversely with wavelength of the radiation.

2.1 Photometry And Radiometry

Photometry is the science of measurement of light, in terms of its perceived brightness to the human eye. It is different from radiometry, which is the science of measurement of light in terms of absolute power. In photometry, the radiant power at each wavelength is weighted by the luminosity function that models human brightness sensitivity. Radiometry is important in astronomy, especially radio astronomy, and is important for remote-sensing applications [11, 12].

2.2 Radiometric Quantities

Radiometry deals with quantities associated with the radiation field, and the interaction coefficients deal with quantities associated with the interaction of radiation and matter. Commonly used radiometric quantities are termed as follows. Radiant flux (φ) is radiant power or energy emitted, reflected , transmitted or received per unit time. The units are Joules per second or watts. The Sun's radiant flux is 3.9×10^{26} W. Monochromatic intensity (I_λ) : The energy transferred by electromagnetic radiation in a specified direction per unit area per unit time normal to the direction considered at a given wavelength is known as monochromatic intensity and given by [12].

$$I = \int (I_\lambda d\lambda) = \int (I_\nu d\nu), \tag{2.2}$$

both I_λ and I_ν are called monochromatic intensity, even though they are expressed in different units. They have the relation as

$$I_\nu = \lambda^2 I_\lambda, \tag{2.3}$$

Monochromatic irradiance (E_λ) is the rate of change of energy transfer per unit area by radiation with given wavelength through a plane surface. Irradiance covers an infinitesimal wavelength interval of the electromagnetic spectrum. The unit used for the measurement of irradiance is Wnm^{-3} [10]. Let the radiation be incident on normal to a horizontal plane, then the monochromatic intensity can be:

$$E_\lambda = \int_{2\pi} I_\lambda \cos(\theta) d\omega, \tag{2.4}$$

where π indicates that the integration extends over the entire hemisphere, θ is the zenith angle, and $d\omega$ represents an element arc of solid angle. The radiant flux divided by the area through which it passes is known as the irradiance(E)and defined as

$$E = \int_0^\infty E_\lambda d\lambda, \tag{2.5}$$

sometimes when referring to the blank's function, $B_\lambda(T)$ may be used in place of E_λ. In general, the irradiance upon an element of surface may consist of contributions which come from infinite different directions [11].

2.3 Black Body Radiation

Black body radiation refers to an object or system which absorbs all radiations incident upon it and reradiates energy which is characteristics of the radiating system only, and not dependent upon the type of radiation which is incident on it. Physically a black body is an ideal perfect radiator and perfect absorber.

All incident radiation falling on the black body is completely absorbed and hence known as black. Maximum possible emission occurs at all wavelengths and in all directions, thus the radiance is independent of direction, known as isotropic radiation [11, 12].

2.3.1 Radiation Laws Governing Black Bodies

All objects will emit radiation when they are given sufficient energy in the form of heat. The emission of this radiation, or electromagnetic radiation, is one of the ways in which objects can lose energy to their lower-temperature surroundings. The radiation is emitted as a continuous spectrum of all frequencies, but the distribution of energy among the frequencies of the spectrum of a black body depends only upon the temperature of the body. The basic laws of radiation governing the relationship between the temperature of the black body and the frequency spectrum of emission are briefly mentioned below [11]

Plank's Law

The emission of radiation of an ideal black body is defined by plank's radiation law. It states that the emitted radiation power increases with increasing temperature and that the maximum of the radiated spectrum is shifted toward shorter wave lengths with rising temperature [13]. Irradiance (E_λ) of radiation from a black body as a function of its temperature T:

$$E_\lambda = B_\lambda(T) = \frac{c_1}{\lambda^5[\frac{exp(c_2)}{\lambda T}] - 1} \tag{2.6}$$

The above equation showing that how radiated energy emitted at shorter wavelengths increases more rapidly with temperature than energy emitted at longer wavelengths. The law may also be expressed in other terms, such as the number of photons emitted at a certain wavelength, or the energy density in a volume of radiation. where c_1 and c_2 are constants, and T is temperature of the black body in k. The shapes of the plank's function for different temperatures are shown in Fig 2.1. It can be seen that the wavelength of the peak of the black body radiation curve decreases in a linear fashion as the temperature increases. The monochromatic intensity shows a sharp cutoff at shorter wavelengths and increases steeply to a maximum. There after, the intensity decreases rather slowly with increasing wavelengths. It explains why hotter bodies attain higher intensity peaks at shorter wavelengths. As the temperature of the black body

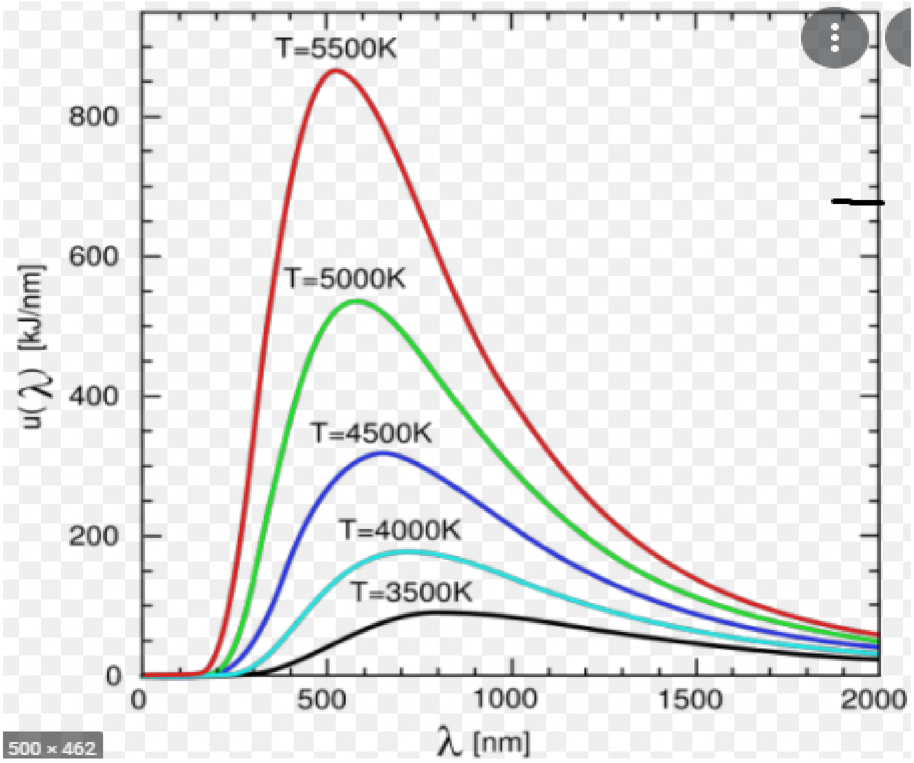

Figure 2.1: Electromagnetic spectrum showing that the temperature is a function of wave length based on the plank's law of radiation [1]

decreases the emission intensity also decreases [1, 14].

Wien's Displacement Law

Wien's displacement law states that there is an inverse relationship between the wavelength of the peak of the emission of a black body and its temperature. Differentiating the plank's function and setting the derivative equal zero gives the wavelength of the peak emission as

$$\lambda_{\max} = \frac{b}{T} \tag{2.7}$$

where $\lambda_{\max}$ is the peak wavelength in meter, b is proportionality constant, called Wien's displacement constant and equals to $2.8978 \times 10^6 nmK$ [1]. Wien's law explains that objects of different temperature emit spectra that peak at different wavelengths. Hotter objects emit most of their radiation at shorter wavelength; hence, they will appear to be bluer. Cooler objects emit most of their radiation at longer wavelengths; hence, they will appear redder. For example, the

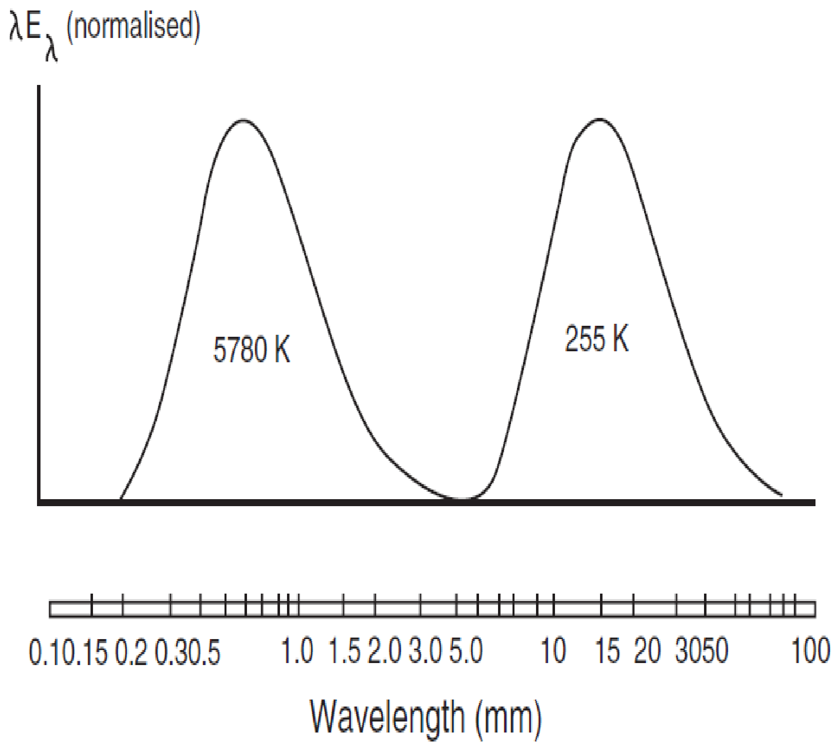

Figure 2.2: Black body spectrum of temperature of the sun and the Earth [2]

surface temperature of the sun is 5788 K, giving a peak at 502 nm, which is fairly in the middle of the visual spectrum. Due to the Rayleigh scattering of the blue light by the atmosphere the white light is separated somewhat, resulting in a blue sky and a yellow sun [1]. Figure (2.2) illustrates the normalized black body spectra of the sun and the earth. Over 99 percent of solar radiation is emitted at wavelengths shorter than 4000 nm. By contrast, radiation emitted from objects such as the Earth lies mainly between 5000nm and 50000nm and peaks around 10000 up to 15000 nm. Further more, at any wavelength, a hotter object is more luminous than a cooler one. According to Wien's displacement law, the wavelength of maximum emission of earth's atmosphere of temperature 255K is 11000nm.

2.3.2 Light from the sun and moon

The surface temperature of the sun is 5778K. Using Wien's law, this temperature corresponding to a peak emission at a wavelength of 2.89777 million nm K/5778K = 502 nm. This wavelength is fairly in the middle of the most sensitive part of land animal visual spectrum acuity. Even nocturnal and twilight hunting animals must sense light from waning day and the moon, which is reflected sun light with this same wavelength distribution. Also, the average wavelength of starlight maximal power is in this region, due to the sun being in the middle of a common temperature range of stars [15].

Raleigh-Jeans Radiation law

The Rayleigh-Jeans law, attempts to describe the spectral intensity(I) of electromagnetic radiation at wavelengths from a black body at a given temperature. For wavelength λ, it is

$$I_\lambda = \frac{2cKT}{\lambda^4}, \tag{2.8}$$

where T is temperature in kelvin, and K is Boltzmann's constant. In the microwave region of the electromagnetic spectrum, when the wavelength is at $5 \times 10^6 nm$ and T is at terrestrial temperature, we find that $hc/\lambda \lll KT$, so the plank's function may be approximated by

$$I_\lambda = \frac{c_1 T}{c_2 \lambda^4}. \tag{2.9}$$

In this rayleigh- Jeans region, the plank monochromatic radiance E_λ is a linear function of temperature T.

Stefan-Boltzmann law

The stefan-Boltzmann law, also known as stefan's law, states that the total radiated per unit surface area of a black body in unit time, E, is directly proportional to the fourth power of the thermodynamic temperature T of the black body [15]

$$E = \sigma T^4. \tag{2.10}$$

The irradiance E has dimension of power density and is measured in Wm^{-2}. T is the absolute temperature and δ is the constant of proportionality known as Stefan-Boltzmann constant, which is non fundamental in the sense that it is an agglomeration of the constants of nature[16]. The value of constant is

$$\delta = \frac{2\pi^5 k^4}{15c^2 h^3} = 5.6704 \times 10^{-8} W m^{-2} K^{-4} \tag{2.11}$$

where k is Boltzmann's constant, h is plank's constant, and c is the speed of light. The surface of the sun is 21 times as hot as that of the Earth, and therefore it emits 190,000 times as much energy per square meter. The distance from the sun to the Earth is 215 times the radius of the sun, reducing the energy per square meter by a factor of 46,000. Taking into account that the cross section of a sphere is one fourth of its surface area, we see that there is equilibrium of approximately $342 W m^{-2}$ surface area, or $1370 W m^{-2}$ cross-sectional area. This shows roughly why the temperature $T \approx 300K$ is approximately that of the Earth. If we assume that the earth is a perfect black body and includes the effect of the Earth's albedo as 30 percent then the Earth's average surface temperature can be estimated as 255 K. But the actual measured value of Earth's surface temperature is 288 K. The difference of 33 K between such calculated value and the actual measured one is due to the effect of greenhouse gases, namely water vapor, carbon dioxide, and methane [17].

2.3.3 Atmospheric Scattering

Scattering is the process by which a particle in the path of an electromagnetic wave obstructs energy from the incident wave and reradiates that energy in all directions. In scattering, no energy transformation takes place, but only a change in a spatial distribution of the energy. Sunlight coming into the atmosphere can be scattered in any direction as it passes through the air. This diffuses the light and spreads it out in all directions so that it does not appear just as a single, straight beam. In the atmosphere, the particles responsible for scattering cover the size from gas molecules $(10^{-8}) cm$ to large raindrops and hail particles of about $1 cm$[17, 18]. The relative intensity of the scattering pattern depends mainly on the ratio of particles size to wavelength of incident wave. There are three different types of scattering: **Rayleigh scattering, Mie scattering and Non-selective scattering**. **Rayleigh scattering**: mainly consists of scattering from atmospheric gases. This occurs when the particles causing the scattering are smaller in size than the wavelengths of radiation in contact with them. This type of scattering is therefore wavelength-dependent. As the wavelength decreases, the amount of scattering increases. Because of Rayleigh scattering, the sky appears blue. This is because of blue light is scattered around four times as much as red light, and UV light is scattered about 16 times as much as red light [19]. **Mie Scattering**:This scattering is caused by pollen, dust, smoke, water droplets, and other particles in the lower portion of the atmosphere. It occurs when the particles causing the scattering are larger than the wavelengths of radiation in contact with them. Mie scattering is responsible for the white appearance of the clouds. **Nonselective scattering**: It occurs in the lower portion of the atmosphere when the particles are much larger than the incident radiation. This type of scattering is not wavelength-dependent and is the primary cause of haze [20].

2.3.4 Absorption And Emission

The maximum amount of radiation that can be emitted by any object at a given temperature is determined by plank's law. A real object, which is not a perfect black body, will emit less than black body radiation. For a given wavelength λ, the emissivity (ε) is defined as the ratio of actual emitted radiance I_λ to the black body radiance B_λ:

$$\varepsilon_\lambda = \frac{I_\lambda}{B_\lambda}, \tag{2.12}$$

The emissivity ranges from 0 to 1 for real objects and is a measure of how strongly a body radiates at a given wavelength. If the emissivity is independent of wavelength, then the emitter is called gray body. Clouds and gases have emissivities which are a strong function of wavelength

. The emissivity of the sea surface is close 1 for visible wavelenghts. The emissivity of a perfect black body is unity, i.e., ε_λ=1, whereas for a gray body it is constant [10]. In the case of selective radiation, the emissivity is a function of wavelength. A high emissivity, near to the maximum of 1, indicates an object that absorbs and radiates a large proportion of the incident energy and a low emissivity(closer to zero) indicates an object that absorbs and radiates a small proportion of the incident energy. The majority of natural objects, excluding water, are selective radiators and their emissivity is wavelength-dependent. In similar manner, we can define the monochromatic absorptivity α_λ, reflectivity r_λ, and transmissivity t_λ as a fraction of the incident monochromatic intensity S that a black body absorbs, reflects, and transmits, respectively, as [10, 11].

$$absorptivity(\alpha_\lambda) = \frac{I_\lambda(absorbed)}{I_\lambda(incident)} \tag{2.13}$$

$$Reflectivity(r_\lambda) = \frac{I_\lambda(reflected)}{I_\lambda(incident)} \tag{2.14}$$

$$transmissivity(t_\lambda) = \frac{I_\lambda(transmitted)}{I_\lambda(incident)} \tag{2.15}$$

Kirchhoff's law

A body in local thermodynamic equilibrium will emit the same amount of energy that it absorbs. Therefore the body does not heat up or cool down. Consider a body which is able to absorb and emit radiation. If I_λ is the incident spectral radiance then the emitted spectral radiance E_λ is

$$E_\lambda = \varepsilon_\lambda = \alpha_\lambda I_\lambda, \tag{2.16}$$

where $\alpha_\lambda I_\lambda$ is the absorbed spectral radiance and α_λ is the absorbance at a given wavelengths. For thermal equilibrium the emitted and absorbed radiation is the same; therefore, in case of a black body:

$$I_\lambda = B_\lambda, \tag{2.17}$$

so

$$E_\lambda = \alpha_\lambda. \tag{2.18}$$

This is known as kirchhoff's law, which means that at a given wavelength, weak absorbers are weak emitters, and conversely, strong absorbers are strong emitters. Kirchhoff's law is a general statement equating, emission and absorption in heated objects, following from general consideration of thermodynamic equilibrium. An object at some nonzero temperature radiates electromagnetic energy. If it is a perfect black body, absorbing all light that strikes, it radiates energy according to the black body radiation formula. More generally, if it is a gray body that radiates with some emissivity multiplied by the black body formula, kirchhoff's law states that at thermal equilibrium, the emissivity of a body equals its absorptivity [12, 13]. The absorptivity or absorbance is the fraction of incident light that is absorbed by the body/surface. In the general form of the theorem, this power must be integrated over all wavelengths and angles. In some cases, however, emissivity and absorption may be defined as depending on wavelength and angel. Kirchhoff's law has a corollary: by conservation of energy, the emissivity/absorptivity cannot exceed 1, so it is not possible to thermally radiate more energy than a black body, at equilibrium. In negative luminescence the angle and wavelength integrated absorption exceeds the material's emission. However, such systems are powered by an external source and are therefore not in thermal equilibrium [13].

2.4 Reflectivity and Transmissivity

Consider a spectral radiance I_λ incident on a slab of an absorbing material which is in local thermodynamic equilibrium(LTE). Figure 2.3 summarizes the radiant energy processes taking place including reflection, absorption, re-emission, and transmission. In Fig .2.3, shown below we can see that the reflected radiation is $r_\lambda I_\lambda$, the transmitted radiation is $t_\lambda I_\lambda$, the absorbed radiation is $\alpha_\lambda I_\lambda$, and the emitted radiation is B_λ which is a function of temperature T [21]. For a black body, closely approximated by the Earth's surface or a dense cloud, $t_\lambda = I_\lambda = 0$, so the absorptivity is 1. For a in LTE conservation of energy means that the absorbed radiation equals the incident radiation minus reflection and transmission contributions, so that:

$$\alpha_\lambda I_\lambda = I_\lambda = r_\lambda I_\lambda - t_\lambda I_\lambda, \tag{2.19}$$

or

$$\alpha_\lambda + r_\lambda + t_\lambda = 1. \tag{2.20}$$

The latter equation says that the processes of absorption, reflection, and transmission account

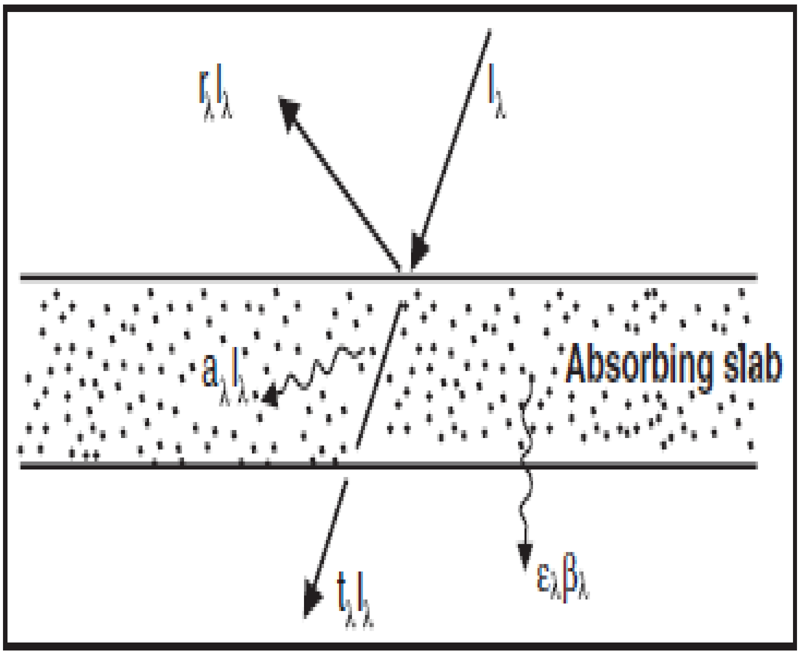

Figure 2.3: Radiative transfer process for monochromatic radiance incident on a slab of absorbing medium [3]

for all the incident radiation falling on a body in LTE. In any atmospheric window region for which

$$t_\lambda = 1, then \;\; \alpha_\lambda = 0 \;\; and, \;\; r_\lambda = 0. \tag{2.21}$$

Monochromatic radiation incident upon an opaque surface with $t_\lambda = 0$ is either absorbed or reflected, so

$$\alpha_\lambda + r_\lambda + t_\lambda = 1 \tag{2.22}$$

At any given wavelength strong reflectors are weak absorbers and weak reflectors are strong absorbers. For the atmosphere, the reflection of IR radiation is negligible, since thee radiation wavelength is large compared to the size of molecules[22]. Therefore, for the atmosphere,

$$\varepsilon = 1 - t_\lambda \tag{2.23}$$

2.4.1 Radiance Temperature

Radiance temperature, sometimes called brightness temperature or black body temperature, is temperature measured by a satellite instrument, usually detected in terms of radiance, but converted into a temperature through plank's function at a given wavelength. The temperature T can be obtained from plank's function as

$$T = \frac{c_2}{\lambda \ln[c_1/\lambda^5 E_\lambda] + 1}, \tag{2.24}$$

2.4.2 Solar Constant

The solar energy reaching the periphery of the Earth's atmosphere is known as the solar constant, and is considered to be constant for all practical purposes. The solar constant is defined as the amount of solar radiation incident, per unit area and time, on a surface which is perpendicular to the radiation and is situated at the outer limit of the atmosphere, the Earth being at its mean distance from the sun. The solar constant consists of all types of solar radiation, and is not just confined to visible light. It is measured by satellite to be roughly $1,366 Wm^{-2}$ [23]. The actual value of the energy varies with several factors, the most important being the Earth's distance from the sun, which changes because of the Earth's elliptical orbit. It fluctuates by amount 6.9 percent during a year from $1,412 Wm^{-2}$ in early January to $1,321 Wm^{-2}$ in early July, due to the Earth's varying distance from the sun, and by a few parts per thousand from day to day. Thus, for the whole Earth, with a cross section of $127,400,00km^2$, the power is 1.740×10^{17} W plus or minus 3.5%. The solar constant is not quite constant over long time periods either. The value $1,366Wm^{-2}$ is equivalent to 1.96 calories per minute per square centimeter [24]. The Earth receives a total amount of radiation determined by its cross section (πR^2), but as the planet rotates, this energy is distributed across the entire surface area ($4\pi R^2$). Hence, the average incoming solar radiation, taking into account the half of the planet not receiving any solar radiation at all, is one fourth the solar constant or $\backsim 342Wm^{-2}$. At any given location and time, the amount received at the surface depends on the state of the atmosphere and the latitude [15, 25].

2.4.3 Albedo

Albedo is known as surface reflectivity of solar radiation. It is defined as the ratio of reflected to incident electromagnetic radiation. Albedo is a measure of surface diffuse reflectivity, which has a non dimensional value, normally represented in percent. The term has its origins from a Latin word *albums*, meaning white. It is quantified as the percentage of all wavelengths reflected by a body or surface to the amount incident upon it. An ideal white body has an albedo of 100%, whereas the albedo is 0% in case of an ideal black body. The albedo of the Earth's surface varies with the type of material that covers it. The typical amounts of solar radiation reflected from various objects are shown in Table 2.4 [15, 4]. Albedo values can range from 3 percent for water at small zenith angles to over 95% for fresh snow. On average the earth and its atmosphere typically reflect about 30%, a number highly dependent on the vegetation, surface characteristics, forest cover, cloud cover and its distribution, etc [26]. Surface reflectance values show large geographic variation. Mean annual albedo values differ considerably between the equator and the poles, largely due to the presence of snow and ice-covered surfaces along with cloudy skies in high latitudes, which greatly increases albedo values in those areas. Atmospheric

Surface	Details	Albedo (in %)
Soil	Wet–dry	4–40
Sand	—	15–45
Grass	Long–short	16–26
Agricultural crops	—	18–25
Tundra	—	18–25
Forests	Desiduous	15–20
	Coniferous	5–15
Water	Small zenith angle	3–10
	Large zenith angle	10–100
Snow	Old/fresh	40–95
Ice	Sea	30–45
	Glacier	20–40
Clouds	Thick	60–90
	Thin	30–50

Figure 2.4: Albedo of various surfaces [4]

reflectance principally varies with dust concentration, the zenith angle of the sun, and the type and/or amount of cloud cover. Well-developed convective clouds reflect up 90% of incident solar energy, making thick clouds appear bright from space. The reflectance properties of the surface change from one season to another. Throughout the high latitudes, snow cover and ice extent reach maximum values during the cold seasons, significantly increasing the surface reflectance values. Melting in the spring exposes bare soils that absorb a significantly greater portion of the incoming solar radiation, decreasing the albedo values [26].

2.4.4 The Greenhouse Effect

The greenhouse effect is the process in which the emission of infrared radiation by the atmosphere warms a planet's surface. The name comes from an analogy with the warming of air inside a greenhouse compared to the air outside the greenhouse. The earth's average surface temperature is about $33c$ warmer than it would be without the greenhouse effect. In addition to the Earth, Mars, and especially Venus have greenhouse effect [27]. A schematic representation of the exchange of energy between outer space, Earth's atmosphere, and its surface is shown in Fig.2.5 indicates. Warming takes place in the atmosphere due to the fact that incoming visible radiation can penetrate to the ground with relatively little absorption, while much of the outgoing long wave infrared radiation is trapped by the atmosphere and emitted back to the ground. In order to satisfy the radiation balance of the Earth atmosphere system, the surface must compensate by emitting more radiation than it would in the absence of such an atmosphere. Thus, greenhouse gases such as water vapor, carbon dioxide, methane, etc. act as an effective blanket around the Earth. The Earth receives energy from the sun in the form of radiation. The earth and its atmosphere reflect about 30% of the incoming solar radiation. The remaining 70% of it is absorbed, warming the land, atmosphere, and oceans. For the earth's temperature to be in steady state, this absorbed solar radiation must be very nearly balanced by energy radiated back to space in the infrared wavelengths. Since the intensity of infrared radiation increases with increasing temperature, one can think of the absorbed solar flux [27]. The visible solar radiation mostly heats surface, not the atmosphere, whereas most of the infrared radiation escaping to space is emitted from the upper atmosphere, not the surface. The infrared photons emitted by the surface are mostly absorbed in the atmosphere by greenhouse gases and clouds and do not escape directly to space. The cause of the surface warming can be easily understood by starting with a simplified model of a purely radiative greenhouse effect that ignores energy transfer in the atmosphere by the convection (sensible heat transport). In this purely radiative case, one can think of the atmosphere as emitting infrared radiation both upward and downward. The upward infrared flux emitted by the surface must balance not only the absorbed solar flux but also this downward infrared flux emitted by the atmosphere. The surface temperature will rise until it generates thermal radiation equivalent to the sum of the incoming solar and infrared radiation. The capacity of the atmosphere to infrared radiation determines the height in the atmosphere from which most of the photons are emitted into space. If the atmosphere is more opaque, the typical photon escaping to space will be emitted from higher from in the atmosphere. Since, the emission of infrared radiation is a function of temperature, it is the temperature at this emission level that is effectively determined by the requirement that the emitted flux balance the absorbed solar flux [5].

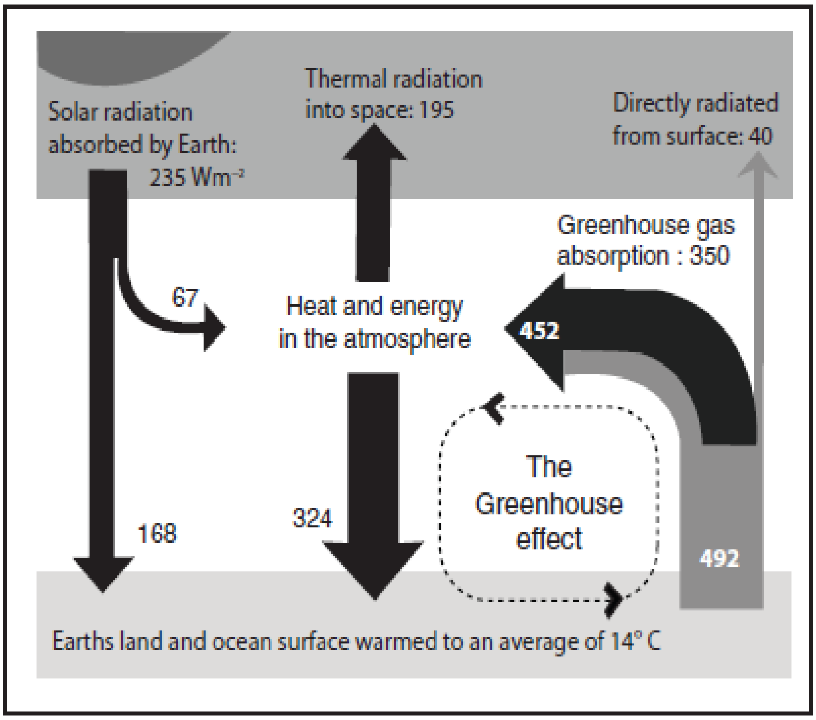

Figure 2.5: Schematic representation of the exchange of energy between outer space and the Erath's atmosphere and its surface [5]

Chapter 3

Radiative Transfer

3.1 Introduction

Both the solar and terrestrial radiations passing through the atmosphere attenuates due to scattering by the aerosols and absorption by gas molecules present in the atmosphere. The attenuation of the radiation depends on: (i) the intensity of radiation at that point, (ii) the local concentration of the gases and aerosol particles that are responsible for the absorption and scattering, and (iii) the effectiveness of the absorbers and scatterers [4].

3.1.1 Beer's law

Beer's law, also known as Bouguer's law or Lambert's law, states that the monochromatic intensity I_λ decreases monotonically with path length as the radiation passes the layer[4, 26]. The law can be demonstrated as follows. Consider a parallel beam of radiation, I_λ, passing through an infinitesimally thin layer of the atmosphere containing absorbing gases and aerosols along a specific path as shown in (Fig. 3.1). After passing the layer, the monochromatic intensity of radiation is decreased by

$$dI_\lambda = -I_\lambda \rho r k_\lambda \sigma ds, \tag{3.1}$$

where ρ is the density of air, r is the mass of the absorbing gas per unit mass of air, and k_λ is the mass absorption or scattering coefficient. As shown in Fig .3.1 the differential path length along the ray path of the incident radiation is ds= $\sec\theta dz$. substituting this value for ds in Equation 3.1, we get

$$dI_\lambda = -I_\lambda \rho r k_\lambda \sec\theta dS. \tag{3.2}$$

Now let us consider the case of depletion of radiation due to absorption and scattering from the

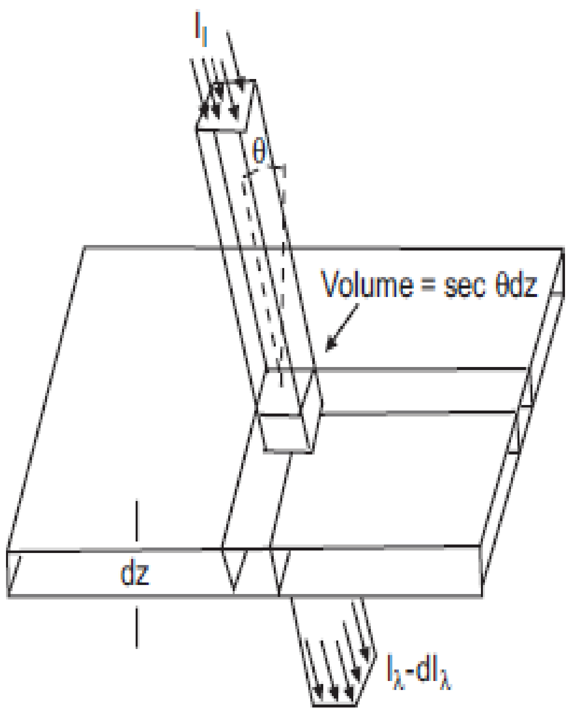

Figure 3.1: Depletion of incoming beam of parallel radiation on passing through a slab of absorbing material [6]

top of the atmosphere (z= ∞) down to any level (z). integrating Eq. (3.2) gives

$$\ln I_{\lambda_\infty} - \ln I_\lambda = \sec\theta \int_z^\infty k_\lambda \rho r dz, \tag{3.3}$$

taking antilog on both sides, we get

$$I_\lambda = I_{\lambda_\infty} \exp(-\tau \sec\theta) = I_{\lambda_\infty} t_\lambda. \tag{3.4}$$

where $\tau_\lambda = \int_z^\infty k\rho$ rdz is a dimensionless quantity referred to as the normal optical depth or optical thickness[28]. It is a measure of the cumulative depletion that a beam of radiation directed straight downward with zero zenith angle would experience in passing through the layer, and $t_\lambda = \exp(-\tau_\lambda \sec\theta)$ is the transmissivity of the layer. In the absence of scattering, the monochromatic absorptivity is

$$\alpha_\lambda = 1 - t_\lambda = 1 - \exp(-\tau_\lambda \sec\theta), \tag{3.5}$$

approaches unity exponentially with increasing optical depth. In the above cases, we have dealt with the scattering and absorption of solar radiation in the atmosphere in the absence of emission. Now we will see the absorption and emission of infrared radiation in the absence of scattering [5, 6].

3.1.2 Schwarszchild's Equation

We can derive the equation that governs the transfer of the infrared radiation through a gaseous medium. The rate of change of the monochromatic intensity of outgoing terrestrial radiation along the path length of ds = $\sec\theta$ dz due to absorption within the layer, is written as

$$dI_\lambda(absorption) = -I_\lambda k_{\alpha\lambda}\rho r \sec\theta dz = -I_\lambda \alpha_\lambda \tag{3.6}$$

where, $k_{\alpha\lambda}$ is the mass absorption coefficient , and α_λ is the absorptivity of the layer. The corresponding rate of change of radiation due to emission is

$$dI_\lambda(emission) = -B_\lambda(T)\varepsilon_\lambda. \tag{3.7}$$

According to Kirchhoff's law, (as seen in Eq. (2.16) the emissivity ε_λ of a black body is equal to its absorptivity of the layer. Then, Equation (3.7) can be written as

$$dI_\lambda(emission) = -B_\lambda(T)k_\lambda \rho r \sec\theta dz \tag{3.8}$$

The attenuation of monochromatic intensity due to both emission and absorption of the layer can be obtained by adding Eq. (3.6) and Eq. (3.8) and we obtain

$$dI_\lambda = -[I_\lambda \alpha_\lambda - B_\lambda(T)]K_\lambda \rho r \sec\theta dz \tag{3.9}$$

This equation is known as Schwarzschild's equation. It states that as the radiation passes through an isothermal layer, its monochromatic intensity exponentially approaches that of a black body radiation corresponding to the temperature of the layer [28].

3.1.3 Solar Absorption And Atmospheric Heating

The absorption of solar radiation in the atmospheric layers can be estimated by using the Beer Lambert law, the intensity of solar radiation of a given wavelength I_λ absorbed by a thin layer of the atmosphere of thickness dz in terms of molecular number density can be represents as:

$$dI_\lambda = -I_\lambda \sigma_{\alpha\lambda} n \sec\theta dz \tag{3.10}$$

where, $\sigma_{\alpha\lambda}$ n= k_α is absorption coefficient, in which $\sigma_{\alpha\lambda}$ is the absorption across section and n is the molecular number density[28]. Integration of this absorption over a path through the atmosphere gives

$$I_\lambda = -I_{o\lambda} \exp[-\int \sigma_{\alpha_\lambda} n(z) \sec\theta dz]. \tag{3.11}$$

The optical thickness is presented as

$$\tau_\alpha = \int \sigma_{\alpha\lambda} n(z) \sec\theta dz. \tag{3.12}$$

The transmission t through the path is

$$t_\lambda(z) = \exp[-\tau_{\alpha\lambda}(z) n(s) \sec\theta ds]. \tag{3.13}$$

Solar radiation penetrates the atmosphere at angle of incidence, which depends on the latitude, season, and local time. By spherical geometry, the cosine of the solar zenith (θ) can be written as

$$\cos\theta = \cos(\phi)\cos(\delta)\cos(HA) + \sin(\phi)\sin(\delta) \tag{3.14}$$

where, ϕ = latitude, δ = solar declination angle 23.5 for solstice, $0 equinox$ or apparent path of the sun, HA = hour angle, for which 0 is local noon. Let us assume a parallel path atmosphere, ds=sec θdz. This approximation is good until we have $\theta > 75$. At that solar zenith angle, we need to take earth's curvature into account. The molecular number density n(z)often varies exponentially with altitude. The number of density represents the concentration of molecules, and at any height z can be represented by the relation

$$n(z) = n_o \exp -(\frac{z}{H}) \tag{3.15}$$

where n_o is the number density at height z=0, and H is the scale height [5, 6]. Thus,

$$\begin{aligned} I(z) &= I_\infty \exp[-(\sec\theta)\int_z^\infty \sigma_a n_o \exp(\frac{-Z}{H})dz] \\ &= I_\infty \exp[-(\sec\theta\sigma_a n_o H)\exp(\frac{-Z}{H})] \end{aligned} \tag{3.16}$$

where, I_∞ is the solar intensity outside of the Earth's atmosphere. The rate of energy disposition by absorption is proportional to the rates of formation of ions, the rates of photo dissociation and production of heat. They are all directly linked to the rate of energy deposition in the atmosphere by absorption. If there is any fluorescence, then the energy deposition will be somewhat less[29].

$$r = \frac{-dI}{sec\theta}dz = \sigma_\alpha n_o I(\infty)\exp[\frac{-z}{H}] + \tau_o \exp[\frac{z}{H}] \tag{3.17}$$

where $\tau_o = \sigma_\alpha n_o H \sec\theta$. The altitude with the maximum absorption, which is found by differentiating the expression r and setting it equal to zero and then solving for z, is :

$$z_m = H\ln(\tau_o z_m) = H\ln(\sigma_\alpha n_o H \sec\theta) \tag{3.18}$$

For an overhead sun, the altitude of maximum absorption is:

$$z_(m) = H\ln(\sigma_\alpha n_o H)z_o \tag{3.19}$$

so that the altitude of maximum energy absorption at any other solar zenith angle is:

$$z_m = z_o + H \ln(\sec\theta) \tag{3.20}$$

The rate of energy deposition at the maximum is given by:

$$r_m = \sigma_a n_o I_\infty \cos\theta \exp[-1 - \frac{Z_o}{H}]. \tag{3.21}$$

The variation of the energy deposition is given by the equation:

$$\frac{r}{r_o} = \exp[1 - Z - \sec\theta \exp(-z)], \tag{3.22}$$

where, $Z = (Z - Z_o/H)$, and r_o is the rate of energy absorption for $\theta = 0$,

$$r_o = \sigma_a n_o I\infty \exp[-1 - \frac{Z_o}{H}]. \tag{3.23}$$

3.1.4 Translation Of Energy Deposition To Heating Rates

If we denote the flux I as units in Wm^{-2}, then the heating rate can be related directly to the rate of energy deposition [30, 7]. The flux of radiation integrated over the entire range of wavelengths can be written as :

$$I = I_\infty \int \exp[-\int \sigma a\lambda n(z) \sec\theta dz] d\lambda. \tag{3.24}$$

Now integrating Eq. (3.24) over a range of wavelengths, only O_2 and O_3 really contribute to the heating:

$$I = \int_\lambda I_\infty \exp[-\int \sigma_{o_2}\lambda[O_2] sec\theta dz - \int \sigma_{o_3}\lambda[O_3] sec\theta dz] d\lambda \tag{3.25}$$

The rate of energy deposition (r) as shown in Eq.(3.17) is

$$r = \int_\lambda I\sigma_{o_2}\lambda[O_2] + \sigma_{o_3}\lambda[O_3] d\lambda \tag{3.26}$$

where r has units of $Js^{-1}m^{-3}$. If all of this energy goes into heating, then the rate of energy deposition equals the rate of heating. By the first law of thermodynamics:

$$\frac{dU}{dt} = Q \tag{3.27}$$

which can be represented as

$$\rho C_p DT/Dt = r \tag{3.28}$$

$$\frac{DT}{Dt} = \frac{1}{\rho C_p}\int_\lambda I\{\sigma_{o_2}\lambda[O_2] + \sigma_{O_3}\lambda[O_3]\}d\lambda \tag{3.29}$$

where ρ is the density, and C_p is the specific heat, or specific enthalpy $(1,005Jkg^{-1}K^{-1})$.

3.1.5 Infrared Heating And Cooling

Infrared heating plays a relatively small role in stratospheric heating. The main infrared heating is by absorption of the 2,700 nm and 4300 nm bands of CO_2. The 96,00 nm band of O_3 provides some of heating near the tropopause. Infrared cooling is primarily due to the 15,000 nm band of CO_2. The second most important infrared cooling is in the 96,00 nm band of O_3, which is most important near the stratopause. We can use the cooling-to-space approximation to estimate the cooling rate. In this approximation, we ignore the downward flux of infrared and look only at the upward flux. This is not a bad approximation, as can be seen in the figure below on upward and downward fluxes [**?**]. The temperature change with time is given by the equation:

$$\frac{DT}{Dt} = -CB_v(T), \tag{3.30}$$

where $B_v(T)$ is the plank's law. C is a constant that depends on the infrared lines, their shape and overlap, and their oscillator strengths [13].

3.1.6 Radiative Heating Due To Absorption

In order to determine the heating rate of the layer between altitudes z and z+dz, the energy balance at each boundary of the layer must be identified. As the thickness dz approaches zero, the energy absorbed per unit volume is given by the net flux divergence $\frac{DT}{Dz}$ [14]. Therefore, the variation of the temperature in the layer per unit time is given by

$$\frac{DT}{Dt} = \frac{-1}{\rho C_p}\frac{DF}{Dz}\frac{DT}{Dt} = \frac{-g}{C_p}\frac{DF}{Dp}, \tag{3.31}$$

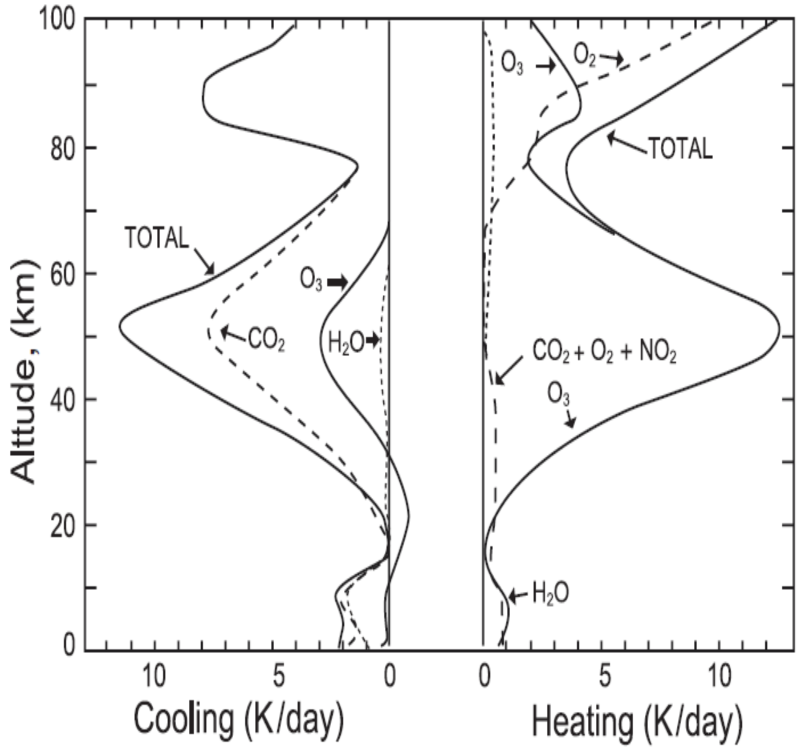

Figure 3.2: Vertical distribution of shortwave heating rates and longwave cooling rates[7]

where, C_p is the specific heat at constant pressure, ρ is the total air density, g is the acceleration due to gravity and p is the pressure.

3.1.7 Vertical Profile Of Radiative Heating

The rate of change of temperature due to absorption or emission of radiation within the atmospheric layer is given by

$$\rho C_p \frac{DT}{Dt} = \frac{DF(z)}{Dz}, \tag{3.32}$$

where F= $F\uparrow + F\downarrow$ is the net flux and ρ is the total density of the air. The rate of heating per unit wave number v is given as

$$[\frac{DT}{Dt}]_v = [\frac{-1}{\rho C_p}\frac{DF(z}{Dz} = \frac{-1}{\rho C_p}\frac{D}{Dz}[\int_{4\pi} I_v \mu d\omega] = \frac{-1}{\rho C_p}\int_{4\pi}\frac{D}{Dz}I_v d\omega = -2\pi\rho C_p\int_{-1}^{1}\frac{D}{Dz}I_v d\mu \tag{3.33}$$

where, $\mu = \cos\theta$, ω is the solid angle, and $ds = dz/\mu = \sec\theta dz$ Substituting for dI_V from schwarszschild's equation, we get

$$[\frac{DT}{Dt}]_v = \frac{2\pi}{\rho C_p}\int_{-1}^{1}[k_v I_v - B_v]d\mu. \tag{3.34}$$

This equation is generally used in estimating the infrared radiative heating rates. The the most important radiatively active greenhouse gases in the atmosphere are carbon dioxide, water vapor, and ozone. In the troposphere, all the three constituents produce radiative cooling in the long wave part of the spectrum. Water vapor is the dominant contributor, but decreases with height. Upper troposphere experiences a net radiative cooling due to the presence of greenhouse gases. On the contrary, the stratosphere is nearly in radiative equilibrium state. Radiative heating due to the absorption of solar radiation in the ultraviolet part of the ozone molecules exactly balances the long wave cooling to space by carbon dioxide, water vapor, and ozone in which carbon dioxide is the most significant contributor to the long wave cooling in the atmosphere [14, 12].

Chapter 4

Solar Radiation And Earth's Atmosphere

Solar radiation is the dominant direct energy source from the stratospheric ozone photochemistry, global atmospheric circulation, tropospheric weather systems and affects all physical, chemical and biological processes. The sun provides a natural influence on the Earth's atmosphere and climate. In order to understand stratosphere-troposphere interaction process, the changes in solar radiation on passing through the earth's atmosphere must be first understood [5].

4.1 Absorption Of Solar Radiation

The vast amount of energy continuously emitted by the sun is dispersed into our space in all directions. Only a small fraction of this energy is intercepted by the Earth and other solar planets. Solar radiation occurs over a wide range of wavelengths. With a surface temperature of 5,780 K, the energy flux at the surface of the sun is approximately $63 \times 10^{6} Wm^{-}2$ [6] . The production of the radiation by the sun depends on the physical and chemical characteristic of the solar atmosphere. However, the energy of solar radiation is not divided evenly over all wavelength regions but rather sharply centered on the wavelength band of 200-2000 nm. In passing through outer space, which is characterized by vacuum, the different types of solar energy remain intact and are not modified until the radiation reaches the top of the earth' atmosphere [28]. In outer space, gamma ray, x-ray, ultraviolet, and infrared radiations are present. Table. (4.1) illustrates attenuation of solar radiation at its various spectral regions due to the absorption of atmospheric constituents. Electromagnetic radiation coming from the sun travels at a speed of $3 \times 10^{8} ms^{-}1$. This radiation comprises wavelengths that vary from the very short gamma rays to the very long microwaves. About 43 percent of the total radiant energy emitted from the sun is in the visible parts of the spectrum. The bulk of the remainder lies in the near-infrared 49% and ultraviolet 7% sections. Less than 1% of solar radiation is emitted as x-rays, gamma waves, and radio waves [30].

Band	Wavelength (nm)	Atmospheric effects
Gamma ray	<0.03	Completely absorbed by the upper atmosphere
X-ray	0.03–3	Completely absorbed by the upper atmosphere
Ultraviolet (UV)	3–300	
UV_c	200–280	Completely absorbed by oxygen, nitrogen, and ozone in the upper atmosphere
UV_b	280–320	Mainly absorbed by ozone in the lower stratosphere
UV_a	320–400	Transmitted through the atmosphere, but atmospheric scattering is severe
Visible	400–700	Transmitted through the atmosphere, with moderate scattering of the shorter waves
Infrared (IR)	700–14,000	
Reflected IR	700–3000	Mostly reflected radiation
Thermal IR	3,000–14,000	Absorption at specific wavelengths by carbon dioxide, ozone, and water vapor, with two major atmospheric windows

Figure 4.1: Solar radiation and its absorption in the Earth's atmosphere [7]

In addition to gamma rays and x-rays, which are absorbed high in the atmosphere, ultraviolet(UV) radiation in the atmosphere is divided into three spectra: UV_a, UV_b, and UV_c as shown in Fig. (4.1). UV_a falls right below visible light, with wavelengths that vary from 320 to 400 nm. Although it is not absorbed by ozone, UV_a is the least energetic and the least damaging of all UV radiation. UV_b radiation which ranges in wavelength from 280 to 320 nm, is more energetic than UV_a and is thought to be harmful to the biosphere. It exists in lesser amounts and is largely absorbed by ozone. UV_c, at 200 to 280 nm which is the most energetic and most damaging but least prevalent of the UV radiation types, is totally absorbed by ozone and normal diatomic oxygen high in the atmosphere. Ozone is the most effective in absorbing radiation at the 250 nm wavelength [5]. In fact, it is 100 times more effective at 250 nm than it is at 350 nm. After ozone absorbs this shortwave radiation, it reradiates at generally longer wavelengths which initially goes in all directions. Some part of the radiation is reabsorbed by other atmospheric constituents. Some make it to Earth's surface, and some return to space. Many factors affect the amount of UV radiation that reaches Earth's surface [30]. In addition to the amount of ozone in the stratosphere, the angle of the sun, length of daylight hours, and path length of radiation through the atmosphere are all determined by latitude and time of year. Solar output, the type and thickness of clouds are also important factors. As the sun's energy spreads through space its spectral characteristics do not change because space contains almost no interfering matter [22]. However, the energy flux drops monochromatically as the same total radiated energy spreads over the surface. As the radiation reaches the outer limit of the Earth's atmosphere, the radiative flux is approximately $1,360Wm^{-2}$. Solar radiation plays two important roles in the atmosphere. One is the heating and cooling of the atmosphere. The infrared part of the solar spectrum can heat and cool the atmospheric layers, whereas the UV/visible part will heat certain layers of the atmosphere depending upon its absorption characteristics. Generally, heating is by solar UV absorption by O_2 in the mesosphere and thermosphere and CO_2(15,000 nm), O_3 (96,000 nm)and H_{2O} 80,000 nm in the lower and middle atmosphere. Photochemical dissociation is the other significant effect of the solar radiation on the atmosphere, caused mainly due to the stratospheric ozone and ions in the ionospheric region. Even though only less than 1 percent of the solar radiation is in the UV, it is responsible for most of the photochemical processes, such as ionization, dissociation, etc. taking place in the Earth's atmosphere [12, 23].

4.2 Atmospheric Window

The atmospheric window refers to the region of electromagnetic spectrum that is relatively free from the effects of atmospheric attenuation. The atmospheric window lies approximately at wavelengths of infrared radiation between 8000 and 14000 nm [29]. The absorption of the principal natural greenhouse gases are concentrated in two belts. Gases such as carbon dioxide and methane, along with less abundant hydro-carbons, absorb at wavelengths longer than the atmospheric window region due to the presence of relatively long C-H and carbonyl bonds. The bonds of water vapor and ammonia absorb at wavelengths shorter than 8000 nm. Except for the bonds in ozone, no bonds between carbon, hydrogen, oxygen, and nitrogen atoms absorb between these two ranges. This means that radiation in the atmospheric window is almost completely reflected from the Earth into space. Without this window, the Earth would become much too warm to support life, and possibly so warm that it would lose its water as Venus did early in solar system history. Thus, the existence of a window in the electromagnetic spectrum is critical to Earth remaining a habitable planet [30]. In recent decades, the existence of the atmospheric window has become threatened by the development of highly unreactive gases containing bonds between fluorine and either carbon or sulphur. The stretching frequencies of bonds between fluorine and other light nonmetals are such that strong absorption in the atmospheric window will always be characteristic of compounds containing such bonds. Bonds to other halogens also absorb in the atmospheric window, though much less strongly. Moreover, the unreactive nature of such compounds that makes them so valuable for many industrial purposes means that they are not removable in the natural circulation of the Earths atmosphere [7].

It is estimated, for instance, that perfluorocarbons (CF_4, C_2F_6, C_3F_8) can stay in the atmosphere for over 50,000 years, which may be an underestimate given the absence of natural sources of these gases. This means that such compounds have an enormous global warming potential. One kilogram of sulphur hexafluoride will, for example, cause as much warming as 23 tons of carbon dioxide over 100 years. Perfluorocarbons are similar in this respect, and even carbon tetrachloride (CCl_4) has a global warming potential of 1,800 compared to carbon dioxide [10].

4.3 Attenuation Of Solar Radiation In The Stratosphere And Troposphere

The penetration of solar radiation into the Earths lower atmosphere depends on the absorption of each constituent present in the stratosphere and troposphere. The stratosphere controls the amount of solar radiation reaching the surface of the Earth, and the troposphere regulates the amount of radiation escaping from the Earths surface into space [11, 12]. The absorption bands of O_3, O_2 (rich in the stratosphere), and O_2, H_2O, and CO_2 (abundant in the troposphere) are critical to maintaining the radiation balance of the Earths atmosphere. Molecular oxygen absorbs ultraviolet as well as some visible and infrared light. Water vapor (H_2O) in the troposphere absorbs highly in the range of 400900 nm and again above 1,200 nm. The other important greenhouse gas, CO_2, has high absorption around 1,400 nm and above. When molecules absorb energy, the absorbed energy may cause a chemical change or it may be re-emitted. Often molecules re-emit energy at wavelengths longer than that at which it was absorbed. Thus when molecules such as H_2O and CO_2 absorb visible or infrared light, they often re-emit it as longer wavelength infrared [11]. Solar radiation emanating from the sun is attenuated, before reaching the ground by the earth's atmosphere, which can be classified into two broad types:

Wavelength (nm)	Absorbtion band	Absorbtion characteristics
175–200	O_2 Schumann Runge bands	Absorption by mesospheric and upper stratospheric oxygen molecules; ozone effects important only in the stratosphere
200–242	O_2 Herzberg continuum O_3 Hartley band	Absorption by stratospheric oxygen molecules, absorption by stratospheric ozone
242–310	O_3 Hartley band	Absorption by stratospheric ozone, leading to $O(^1D)$
310–400	O_3 Huggins bands	Absorption by stratospheric and tropospheric ozone, leading to $O(^3P)$
400–850	O_3 Chappuis bands	Absorption by tropospheric ozone causes photodissociation even at the surface

Figure 4.2: Significant absorption bands of solar radiation in the stratosphere and troposphere [1]

(1) atmosphere without clouds and (2) atmosphere with clouds. Maximum radiation on the earth is received under cloudless and clear skies [11]. Most solar devices operate when radiation is at a maximum or at least above a certain threshold level. However, high level of radiation can create serious problems in, for instance agriculture and architecture. Consequently, the cloudless condition is important from both the utilization and the control point of view. In the stratosphere, the energy balance shows that the dissociative heating due to absorption of ultraviolet radiation more than balances the energy lost during destruction and radiation. Hence there is a net warming of the atmosphere in the upper layers of the stratosphere, though ozone is

concentrated in the lower layers. In general the ozone content in the stratosphere is much greater in high latitudes than that in low latitudes. Also, in higher latitudes, there is a large annual variation with very large values in spring as compared to those in autumn. In the equatorial stratosphere, not only is the ozone content small but it has also a small annual variation [2]. Table (4.2) illustrates significant absorption bands of solar radiation in the stratosphere and troposphere. Solar radiation below 100 nm is almost completely absorbed above 100 km by molecular and atomic oxygen and molecular nitrogen. At wavelengths greater than 100 nm, solar ultraviolet radiation can photo dissociate atmospheric molecules. The solar $Lyman-$ line at 121.6 nm is very intense and can penetrate into the upper part of the middle atmosphere, where it effectively dissociates water vapor, carbon dioxide, and methane. At longer wavelengths, the solar spectrum is subdivided into regions of absorption by the principal absorbing species, O_2 and O_3. For wavelengths in the region of 175200 nm, known as O_2 Schumann Runge band, the radiation is absorbed by the O_2 molecules present in the mesosphere and upper stratosphere[31]. In this region absorption by ozone takes place only in the stratosphere. In the 200 up to 242 nm band region, there are two strong absorption bands present Absorption by stratospheric oxygen takes place in the O_2 *Herzberg continuum* and the absorption by stratospheric ozone molecules in the O_3 *Hartley band*. The ozone molecules which are abundant in the stratosphere absorbs almost all radiation in the 242-310 nm spectral region. In the O_3 *Hungginsbands*(310-400), the radiation is mainly absorbed in the stratosphere and troposphere. In the visible and IR part of the solar spectrum, i.e., above 400 nm, the greater part of the solar photons reaches the troposphere and surfaces. In this, region, the effects of molecular scattering, clouds, and surface albedo become significant.

4.4 Stratospheric Cooling

Cooling of the stratosphere is favored by the ozone reduction. But the main cause of stratospheric cooling is the release of carbon dioxide in the troposphere. Therefore, global warming (i.e., tropospheric warming) and stratospheric cooling are parallel effects. Further cooling of the stratosphere may have an impact on the future development of the ozone layer, because a cold stratosphere is necessary for ozone depletion. We may therefore keep in mind that releasing more CO_2 may favor the ozone hole formation[8, 11, 12].

4.5 Causes Of Stratospheric Cooling

There are several causes for the stratosphere cooling. The two better known reasons are: (i) the depletion of stratospheric ozone and (ii) the increase in atmospheric carbon dioxide.

Cooling Due To Ozone Depletion

In the case of less ozone in the stratosphere, absorption of solar UV radiation is reduced. As a result, lesser amount of solar energy is transformed into heat in the stratosphere. The stratospheric heating is thereby reduced because of the reduced absorption of solar UV light. Ozone in the lower stratosphere also acts as a greenhouse gas and absorbs infrared radiation. At about 20 km altitude the effects of UV light and infrared radiation effect are nearly equal. The lower stratosphere seems to be cooling by about $0.5C$ per decade [8, 11, 12]. This general trend is interrupted by heavy volcanic eruptions which lead to a temporary warming of the stratosphere for 12 years. Afterward, temperatures again begin the increasing trend. Mechanism of cooling and heating by absorbing ozone in the atmosphere play a major role in the energy and radiation balance of the earth-atmosphere system . Down welling radiation causes heating of the earth's surface due to direct sunlight absorption and also due to the back radiation from the atmosphere, which is the source term of the heavily discussed atmospheric greenhouse or atmospheric heating effect. Upward radiation contributes to cooling and ensures that the absorbed energy from the sun and the terrestrial radiation can be rendered back to space and the earth's temperature can be stabilized. For all these processes, particularly, the interaction of radiation with infrared active molecules strongly absorb terrestrial radiation, emitted from the earth's surface, and they can also be excited by some heat transfer in the atmosphere. The absorbed energy is reradiated uniformly into the full solid angle but to some degree also re-absorbed in the atmosphere, so that the radiation underlies a continuous interaction and modification process over the propagation distance [8, 12]. Stratospheric cooling may have been taking place over recent decades for a number of reasons. one of the reason may be that the presence of ozone itself generates heat, and ozone depletion cools the stratosphere. Another contributing factor to the cooling may be that rising amount of greenhouse gases in the lower atmosphere (troposphere) are retaining heat that would normally warm the stratosphere [18].

Cooling Due To The Greenhouse Effect

Greenhouse gases CO_2, O_3, CFC generally absorb and emit the infrared heat radiation at a certain wavelength. If this absorption is very strong as in the 15,000 nm absorption band of CO_2, the greenhouse gas can block most of the outgoing infrared radiation already close to the Earth's surface. Nearly no radiation from the surface can, therefore, reach the CO_2 residing in the upper troposphere or lower stratosphere. On the other hand, CO_2 emits heat radiation to space. In the stratosphere this emission becomes larger than the energy received from below by absorption. In total, CO_2 in the lower stratosphere and upper troposphere loses energy to space. It cools these regions of the atmosphere [13]. Other greenhouse gases, such as ozone and chlorofluorocarbons (CFCs), have a weaker impact, because their absorption in the troposphere is smaller. They do not entirely block the radiation from the ground in their wavelength regimes and can still absorb energy in the stratosphere and heat this region of the atmosphere. Figure (4.1) shows the time series of global mean temperature anomalies in the lower stratosphere

(1621 km), tropopause layer (9.516 km), mid-troposphere (1.59.5 km), and surface. The years of major volcanic eruptions are indicated at the bottom of the figure. The figure shows that the stratosphere has cooled significantly since about 1980, mostly as a result of ozone loss but also because of the accumulation of greenhouse gases in the troposphere. Prior to 1980, the tropopause shows a warming trend, which turns into a cooling trend in the later years. In both the surface and mid-troposphere levels, cooling was observed between 1962 and 1976, except for the year 1973. Afterward, a warming trend is generally observed in these layers [13, 18].

4.5.1 Stratospheric Cooling Rates

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. Figure (4.2) shows how water vapor, carbon dioxide, and ozone contribute to long wave cooling in the stratosphere. Especially for CO_2 it is obvious, that there is no cooling in the troposphere, but a strong cooling effect in the stratosphere. Ozone, on the other hand, cools the upper stratosphere, but warms the lower stratosphere [7].

Stratospheric Ozone

Stratospheric ozone is the most important minor constituent present in the Earths 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 [30]. This property of the ozone protects the life on Earth by preventing the harmful radiation reaching the Earths surface. 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 Earths atmosphere. Stratospheric ozone is considered good for humans and other life forms because it absorbs UV_b radiation from the Sun. Otherwise, UV_b would reach Earths surface in amounts that are harmful to a variety of life forms. 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 [18, 30].

The impact of ozone decrease is more relevant in the lower stratosphere in the region around 20 km of altitude. The impact of cooling due to increase of carbon dioxide is high in the upper stratosphere between 40 and 50 km (see Fig.4.2). Due to these different effects, cooling is not homogeneous over the whole stratosphere.

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. About 99 percent of the Earths 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 [9].

Water Vapor

Water vapor is unique among atmospheric trace constituents in that conditions for saturation are common in the atmosphere. 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 percent or 4 percent by volume. Water vapor content of air is a strong function of air temperature. 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 6km where a high proportion of moisture lies [15].

Chapter 5

Summary And Conclusion

Over all this project gives the following conclusions;

Radiation is the process through which energy moves through the space from the source without material medium. In stratosphere the net heating depends only on the imbalance between local absorption of solar UV radiation and infrared radiative loss. The most common Electromagnetic spectrums are: radio wave, micro wave, infrared wave, visible wave, ultraviolet wave, x-ray and gamma ray. The emission of radiation or Electromagnetic radiation is one of the ways in which objects can lose energy to their lower temperature surroundings. Atmospheric scattering is the process by which particles reradiate the electromagnetic radiation they absorb. The radiative process constitute the ultimate source and sink of energy in the atmospheric system. The temperature patterns in the troposphere and stratosphere are the result of difference in the way radiative energy is transferred to the atmosphere. The maximum amount of radiation that can be emitted by any object at a given temperature is determined by planks law. For a given wavelength λ, the emissivity (ε) is defined as the ratio of actual emitted radiance I_λ to the black body radiance B_λ:

$$\varepsilon_\lambda = \frac{I_\lambda}{B_\lambda} \tag{5.1}$$

The solar energy reaching the periphery of the Earth's atmosphere is known as the solar constant, and is considered to be constant for all practical purposes. The solar constant is defined as the amount of solar radiation incident, per unit area and time, on a surface which is perpendicular to the radiation and is situated at the outer limit of the atmosphere, the Earth being at its mean distance from the sun. The solar constant consists of all types of solar radiation, and is not just confined to visible light. It is measured by satellite to be roughly $1,366 Wm^{-2}$. The penetration of solar radiation into the Earths lower atmosphere depends on the absorption of each constituent present in the stratosphere and troposphere. The stratosphere controls the amount of solar radiation reaching the surface of the Earth, and the troposphere regulates the amount of radiation escaping from the Earths surface into space. The absorption bands of O_3, O_2 (rich in the stratosphere), and O_2, H_2O, and CO_2 (abundant in the troposphere) are critical to maintaining the radiation balance of the Earths atmosphere. Cooling of the stratosphere is favored by the ozone reduction. But the main cause of stratospheric cooling is the release of carbon dioxide in the troposphere. Therefore, global warming (i.e., tropospheric warming) and stratospheric cooling are parallel effects. Further cooling of the stratosphere may have an impact on the future development of the ozone layer, because a cold stratosphere is necessary for ozone depletion. We may therefore keep in mind that releasing more CO_2 may favor the ozone hole formation.

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