Chapter 1

Introduction

1.1 Definition of Environmental Fluid Dynamics

Environmental fluid dynamics concerns the flow of fluid in the environment and its use and management.

Human activity is integrated into the environment in which we live. Our health and comfort is determined by the flow and quality of the air we breathe and the water we drink. We harness air and water flows for power and discharge waste into the atmosphere, oceans and rivers. In order to continue sustainable economic activity it is necessary to manage these flows to maintain their quality.

Much of our environment is fluid; obviously the atmosphere, oceans, lakes and rivers. Indeed on longer time scales even the 'solid' earth can be considered as fluid, responsible for the motion of the continental plates (figure 1.1) and the formation of mountains.[1]

Our concerns also span wide ranges of spatial and temporal scales. From short term, local concerns like the proximity of passive cigarette smoke, to long term, wide ranging flows that influence climate change. And, since we live on the earth, environmental fluid dynamics has much in common and significant overlap with geophysical fluid dynamics, which is concerned with the dynamics of the oceans and atmosphere.

For the purposes of this course, we restrict our definition of environmental fluid dynamics to those flows where humans provide some kind of engineering role. This could be the management of a lake as a supply of drinking water, or the harnessing of the wind for wind power or to ventilate a building. The distinction from geophysical fluid dynamics is not precise, and the edges are

[1] Alfred Wegener (1880-1930), a German meteorologist and geologist, was the first person to propose the theory of continental drift. In his book, "Origin of Continents and Oceans," he calculated that 200 million years ago the continents were originally joined together, forming a large supercontinent. He named this supercontinent Pangaea, meaning "All-earth". He was also known for his papers on lunar craters, which he correctly believed were the result of impacts rather than vulcanism.

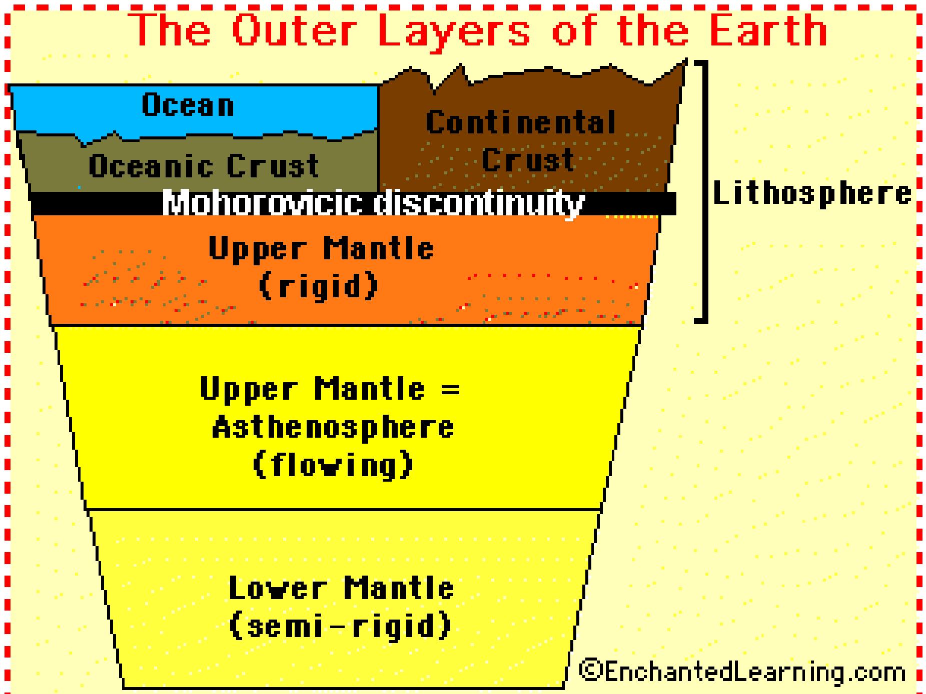

Figure 1.1: Schematic of the outer layers of the earth. From www.enchantedlearning.com

blurred. But since we can not, as yet, change the weather or the flow of the Gulf Stream, we consider the dynamics of these and similar flows to be a topic in geophysical fluid dynamics rather than environmental fluid dynamics.

Engineered flows, such as for example, the flow inside an engine or a gas pipeline, are primarily determined by the geometry - the boundary conditions. Geophysical flows are broadly independent of boundary conditions; instead they are governed by the internal dynamics. Environmental fluid dynamics is concerned with flows that span both aspects - they are natural flows but where the effects if boundaries is usually important (figure 1.5).

For example we will study the sea breeze. Figure 1.6 shows the winds measured at La Jolla during the summer. This figure shows that the wind has a strong diurnal (daily) cycle, with on-shore winds in the afternoon. This sea breeze is driven by the land-sea temperature contrast, and so depends on the local boundary conditions at the coast. It plays a large role in pollutant transport in coastal regions, such as the San Diego - Los Angeles corridor. However, the sea breeze, which is a form of gravity current, could equally well be studied as a geophysical flow, since the flow is really set by the internal dynamics associated with the density changes in the air.

Figure 1.2: Tectonic plates.

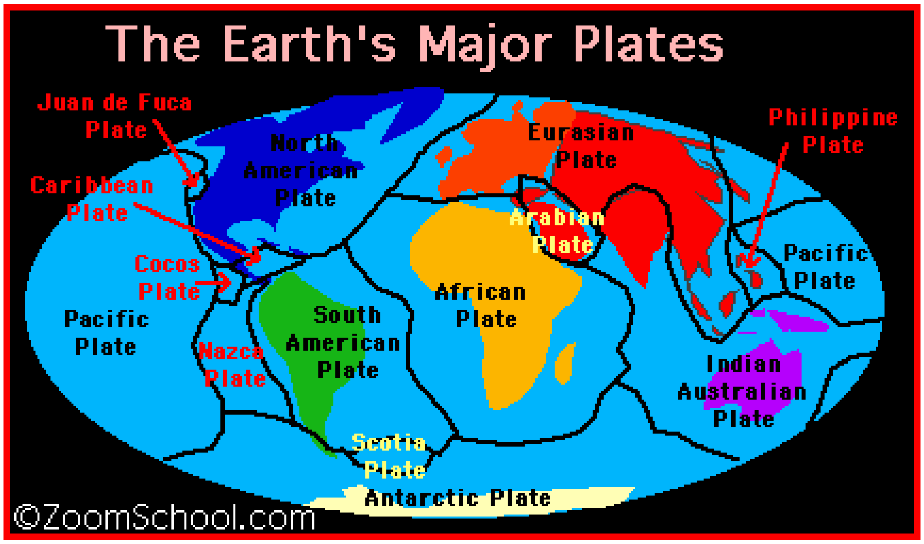

Figure 1.3: The continental plates. From www.enchantedlearning.com

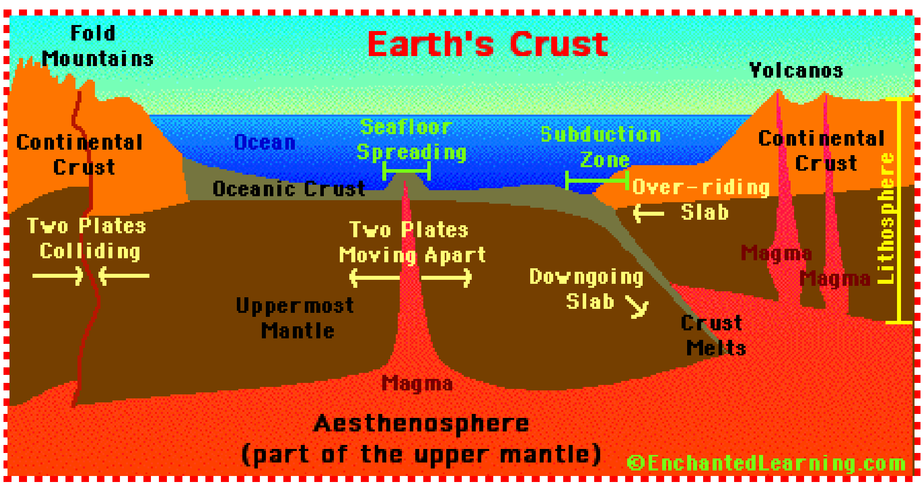

Figure 1.4: Schematic of processes involved in plate tectonics. Note the presence of convective plumes and stable stratification. From www.enchantedlearning.com

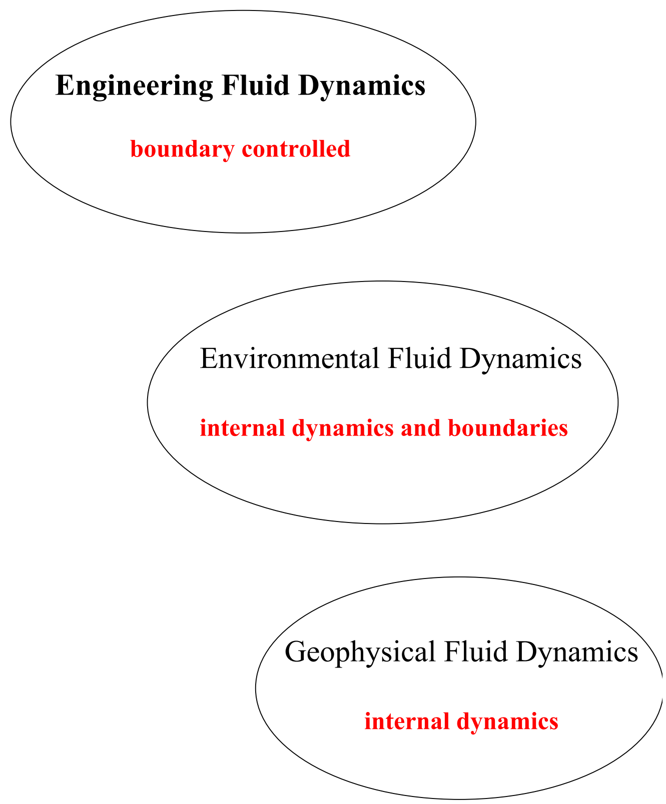

Figure 1.5: Schematic of Environmental Fluid Dynamics.

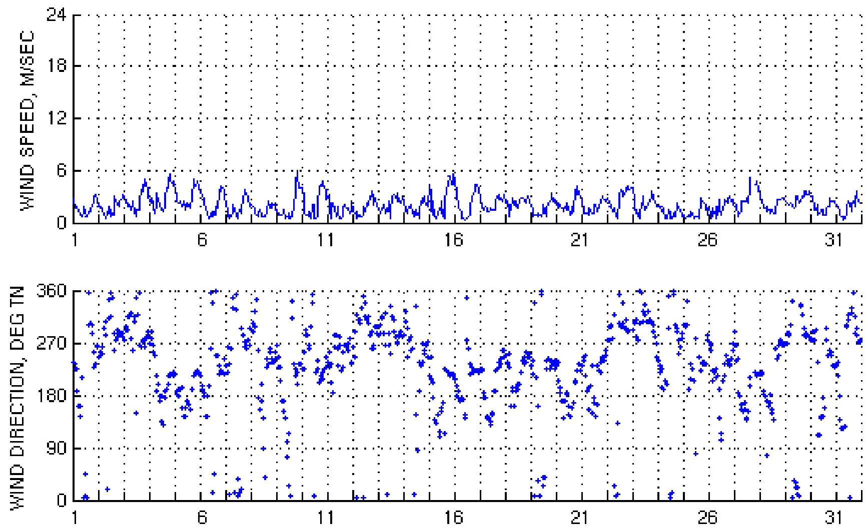

Figure 1.6: Wind speed and direction measured at the Scripps Institution of Oceanography pier in July 2002. Note the clear diurnal cycle with the maximum onshore wind corresponding to the sea breeze.

Environmental engineering is essentially an interdisciplinary subject, since environmental effects include chemical and biological changes. Sustainable engineering also needs to work within the economic and policy framework, possibly changing these to be more environmentally responsible. Similarly, environmental fluid dynamics, particularly the study of the physics of environmental flows, is only one part of the environmental picture. It is, however, the central part, since without an understanding of the physics it is very difficult to remediate or prevent biogeochemical problems in the environment.

As we will see the physics is complex and the equations that govern these flows and their consequences are tough to solve. It is almost always necessary to make approximations both to the real environmental problem and to the equations that describe them in order to get solutions. This process is 'modeling', whereby a problem is analyzed and broken down to its essential parts. Often models, the answers from which are increasingly the basis of decisions costing millions of dollars, are used as black boxes.

It is essential to realize that these models are approximate and have limitations in their applicability and accuracy. The job of the environmental engineer is to be aware of these limitations and to interpret the results in defensible terms. The purpose of this course is to provide some of the background required for this difficult, challenging and important task.

Our environment consists of the atmosphere and the oceans. We begin with

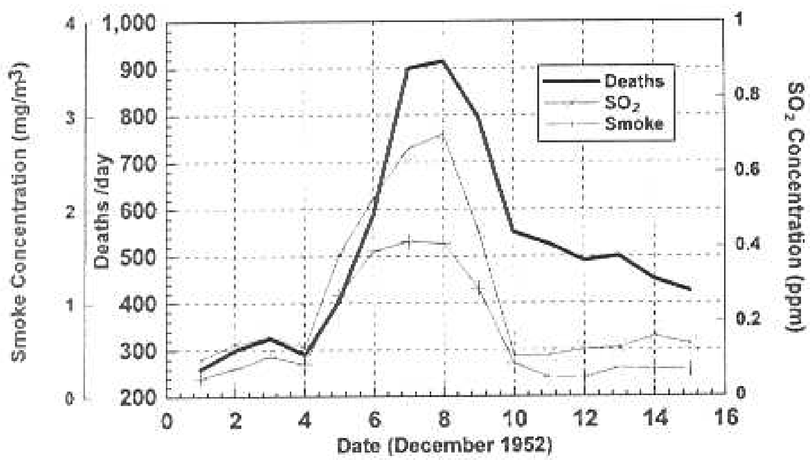

Figure 1.7: Smoke and SO_2 concentrations compared to deaths per day in London during December 1952.

a discussion of the atmosphere, partly because we are all exposed to it and also since it has most of the features that concerns environmental fluid dynamics.

1.2 The atmosphere

We all rely on the atmosphere to provide us with oxygen for life. We have become increasingly aware since the start of the industrial revolution that industry can pollute this air, with deleterious effects on our health, on flora and fauna and even on buildings. Many cities in Europe have been cleaning the grime from their medieval buildings which has built up since the 18^{th} century (figure 1.11). Recent increases in childhood occurrences of asthma in many countries have been attributed to emissions from vehicles. The infamous smog in London in 1952, in which 4000 people died, was the spur for the first clean-air act. This is the first piece of environmental legislation, which has been a model for environmental regulations in an increasing numbers of areas. Figure 1.7 shows the fatalities during the smog, showing that they are correlated *in time* with smoke and SO_2 concentration. A similar correlation was also noted in New York during a pollution episode in 1966 (figure 1.8).

Since that time legislation has caused a significant reduction in the emissions and smoke concentration. Figure 1.9 shows the reduction from the period 1952–1970. This reduction in emissions is mainly due to the introduction of smokeless coal for domestic use. Further reductions since then have mainly been due to reduction in vehicle emissions. It is now possible to predict pollu-

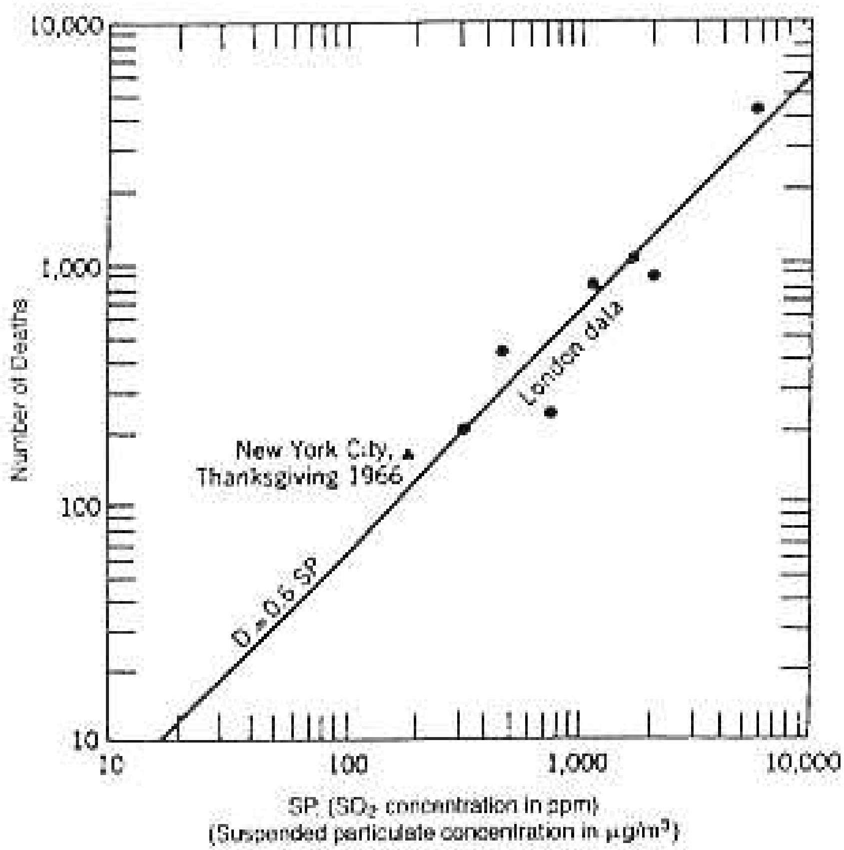

Figure 1.8: Number of deaths in London and New York air pollution episodes as a function of the product of SO_2 and suspended particulate concentrations. From Larsen (1970).

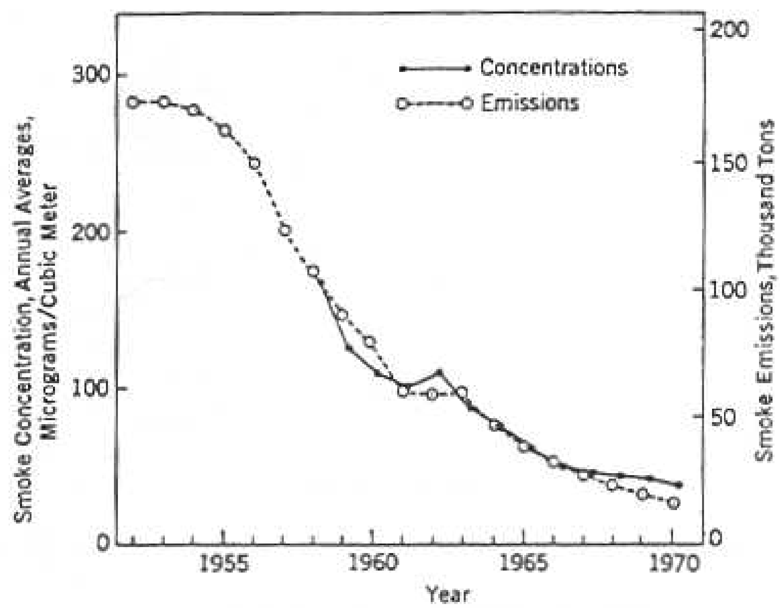

Figure 1.9: Smoke concentration in London from 1952 to 1970. From Hov, Isaksen & Hesstvedt (1976).

tion from emissions with spatial resolution of 10m (figure 1.10).

The atmosphere consists of 3 layers: the surface layer, the troposphere and the stratosphere (figure 1.12). Figure 1.13 shows the 'Standard atmosphere', with the average temperature plotted against height. The vertical scales are height and pressure and density. The density of the air decreases rapidly with height, so that most of the mass of the atmosphere is contained in the lower 10km. Since momentum is proportional to the mass, this implies that the dynamical processes are most significant near the surface and get weaker with height.

Exercise 1.1 *2002 was the 50^{th} anniversary of the 'great London fog'. Look on the web to see what you can find about the fog and what has happened since to mitigate the problem.*

1.2.1 Atmospheric surface layer

This is the lower most part of the atmosphere and it is the most affected by diurnal variations. It extends from the ground to about 1-2 km during the day when the convective heating by the sun is the strongest. At night it is generally

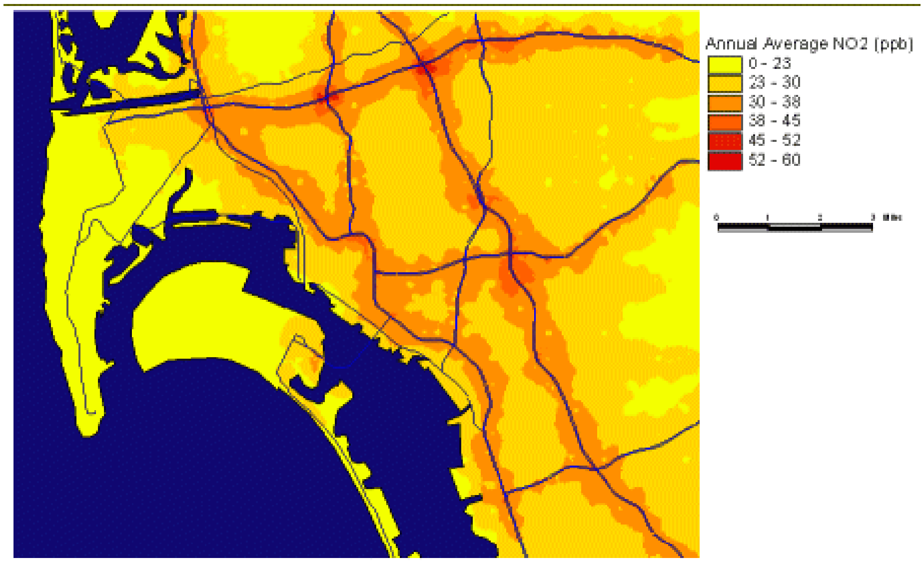

Figure 1.10: Air pollution map of San Diego, obtained using *ADMS Urban*.

much shallower. This region of the atmosphere is where we live and to a large extent is where we release pollutants and extract energy. The wind is turbulent and changes direction with height, the latter due to the rotation of the earth. The temperature is fairly constant with height.

As the day progresses the temperature near the ground increases as a result of the solar heating. This heating causes the air to convect, and mix with the air above. This convection increases the depth of the surface mixed layer during the day. A schematic of this process is shown in figure 1.14. At night (figure 1.14 (a)) the ground cools by long wave radiation and so the temperature increases with height. This stratification is stable (see § 3.2) and, since it is the opposite of the usual situation where the temperature decreases with height, this is called an *inversion*. If the lower part of the atmosphere is stirred by the air flow, the temperature will be uniform up to some height h, and the ground temperature will increase by $\frac{1}{2}hT_z$, where T_z is the initial temperature gradient.

Convection adds extra heat into this lower layer as shown in figure 1.14 (c). If it is assumed that the convection extends up to the height where the ground temperature equals the temperature at $z = h$, then the additional heat that is added is again $\frac{1}{2}hT_z$. We will discuss whether this is realistic in § 3.3.

We will see later on that these changes in the temperature profiles near the ground have a profound effect on the flow in the lower atmosphere, with important implications for the spread of pollutants released from ground sources such as vehicles, industrial plants etc.

Figure 1.11: The Senate House (left) and Caius College, Cambridge in June 2003, showing the effects of air pollution.

1.2.2 Troposphere

This is the region above the ASL to about 10 km. The temperature decreases with height, largely as a result of the decrease in pressure. Most of the mass of the atmosphere is contained below the top of the troposphere, called the tropopause.

In order to determine how the temperature of the air decreases with height in a stationary atmosphere, we need to do a little thermodynamics. Dry air is well represented by the perfect gas equation

$$p = \rho R T, \tag{1.1}$$

where p is the pressure, ρ is the density, T is the temperature (in Kelvin) and R is the gas constant.

We consider what happens when a small volume of air (a parcel) is moved vertically in a stationary atmosphere. The first law of thermodynamics states that the change in heat content dQ is given by

$$dQ = c_V dT + p dV, \tag{1.2}$$

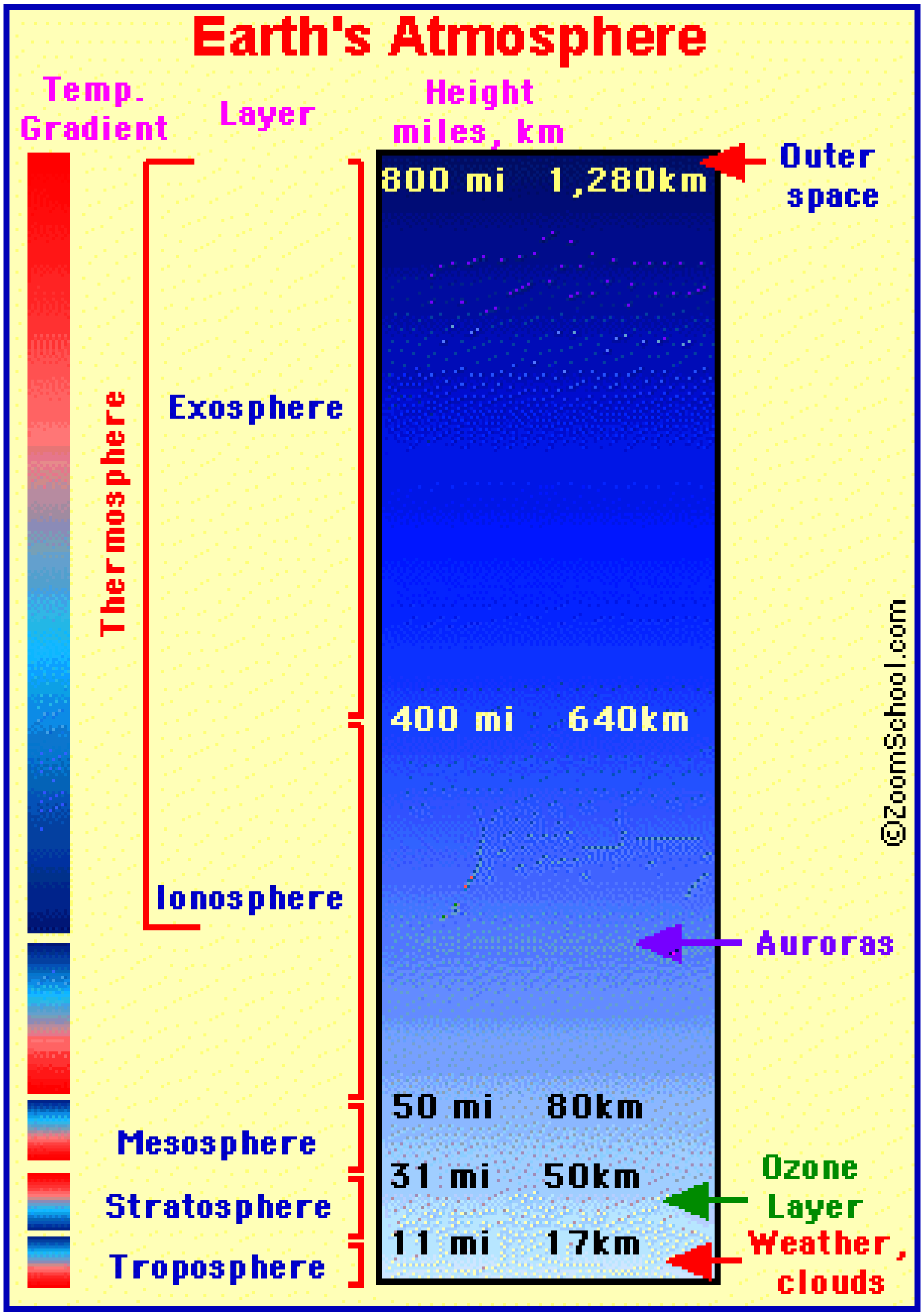

Figure 1.12: The atmosphere. From www.enchantedlearning.com

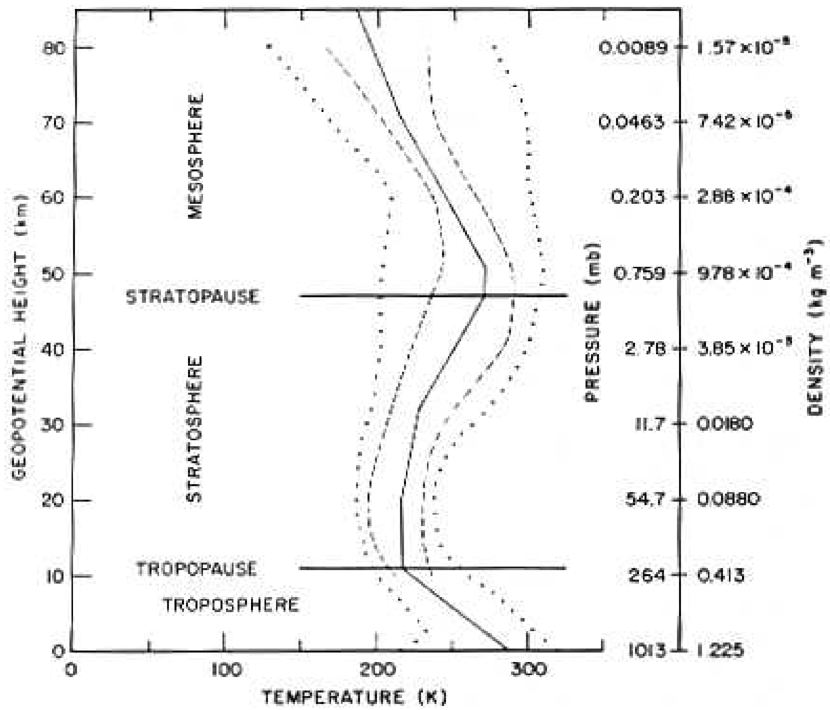

Figure 1.13: Temperature variation with geopotential height of the US Standard Atmosphere (solid line) This consists of straight line segments with breaks at 11, 20, 32, 47, 51 and 71 km. The surface temperature is 15°C and the gradients, starting from the surface, are -6.5, 0, 1.0, 2.8, 0, -2.8 and -2.0 K km^{-1}, respectively. The dashed line shows the lowest and highest monthly mean temperatures obtained for any location between the equator and the pole, while the dotted lines shows estimates of the 1% maximum and minimum temperatures that occur during the warmest and coldest months, respectively, in the most extreme locations. The scale at the right gives the pressures and densities at 10km intervals for the standard profile. From NOAA/NASA/USAF, 1976.

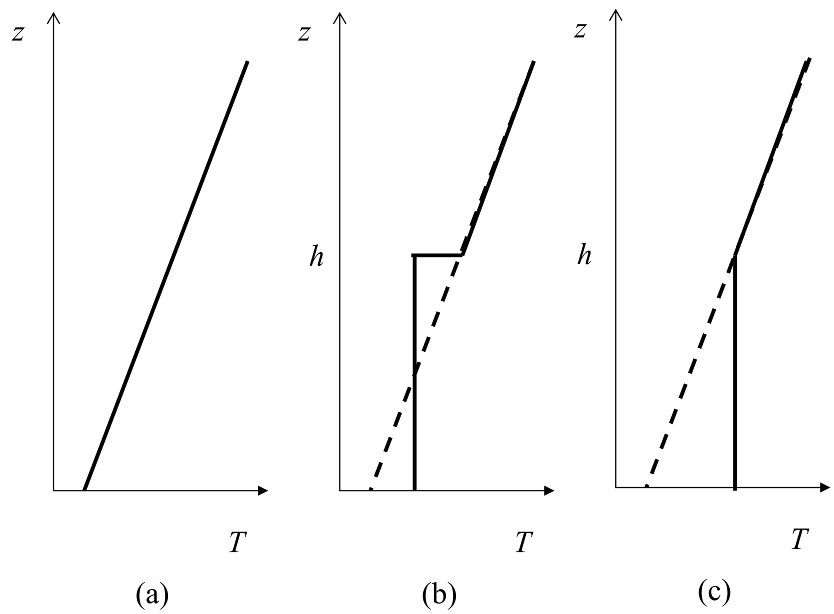

Figure 1.14: Schematic of development of the temperature during the day. (a) Night time conditions, where the temperature is coolest near the ground due to longwave radiation. This is known as an 'inversion'. (b) Redistribution of the temperature caused simply by mixing fluid from below without any heat input. (c) The development when there is heat input and mixing.

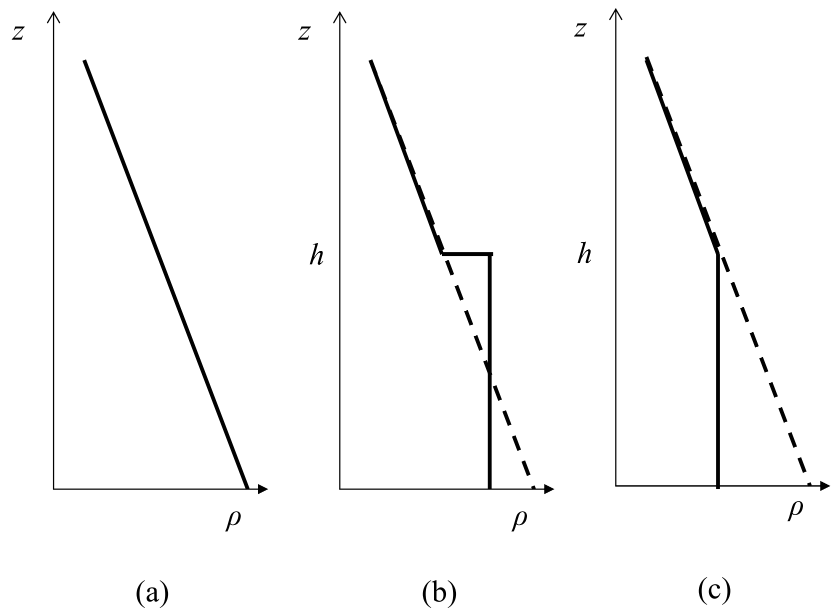

Figure 1.15: Schematic of development of the density profile during the day corresponding to the temperature variation shown in figure 1.14.

where c_V is the specific heat at constant volume and dV is the volume change of the parcel. For an *adiabatic* change, in which no heat is exchanged between the parcel and its surroundings, $dQ = 0$, and writing $V = \frac{1}{\rho}$, and using the equation of state (3.10) we have

$$dp = (c_V + R)dT = c_p dT, \tag{1.3}$$

where $c_p = c_V + R$ is the specific heat at constant pressure.
For a fluid at rest, the pressure (force per unit area) is the weight (per unit area) of the fluid above, i.e.

$$p = \int \rho g dz, \tag{1.4}$$

where the integral is from the point in question to the top of the fluid. This equation is usually expressed as a differential equation

$$\frac{dp}{dz} = -g\rho(z), \tag{1.5}$$

where the negative sign results from the fact that z is measured upwards and the pressure increases downwards. The equation (1.5) is known as the *hydrostatic equation*.

From (1.3) we have

$$\frac{dp}{dT} = c_p. \tag{1.6}$$

Then (1.5) gives

$$\frac{dT}{dz} = -\frac{g}{c_p}, \tag{1.7}$$

and the right hand side of (1.7) is known as the *adiabatic lapse rate*. This means that if a parcel of air is raised adiabatically its temperature will decrease at the rate given by (1.7). For air this rate is about 10 K km^{-1}.

Figure 1.13 shows that the observed temperature decrease is very close this value. Consequently, this implies that the troposphere is close to thermodynamic equilibrium, in that the temperature distribution is close to that achieved simply by having the air at rest. We will return to this point later.

1.2.3 Stratosphere

The stratosphere is the region above the tropopause, i.e above 10 km, and it is generally marked by an increase in temperature with height. As a result this region of the atmosphere is stably stratified (see § 3.2) and there is little vertical mixing. This is one reason why the ozone hole (figure **??**) that occurs in the stratospheric winter hemisphere and the anthropogenic CFCs that contribute to its formation persist for a long time.

Exercise 1.2 *Find out the scale and persistence of the ozone hole. Why does it vary from year to year? What is a Dobson unit?*

1.3 Climate change

The begin of industrialization was the industrial revolution which began in England in the 18^{th} century. It was largely driven by technological innovations such as the use of steam[2], iron and the organization of labor. Since then industrial activity has spread around the globe and increases continually. This activity uses natural resources and produces waste products. Such activity is only sustainable if it does not deplete resources or create waste disposal problems to the extent that the activity can no longer continue.

1.3.1 Radiation balance

The energy for all life and motion on the earth is received as radiation from the sun. Figure 1.19 shows the radiation balance of the earth. The average energy flux from the sun at the mean radius of the earth is called the solar constant S

[2] James Watt (1736–1819) invented the steam engine.

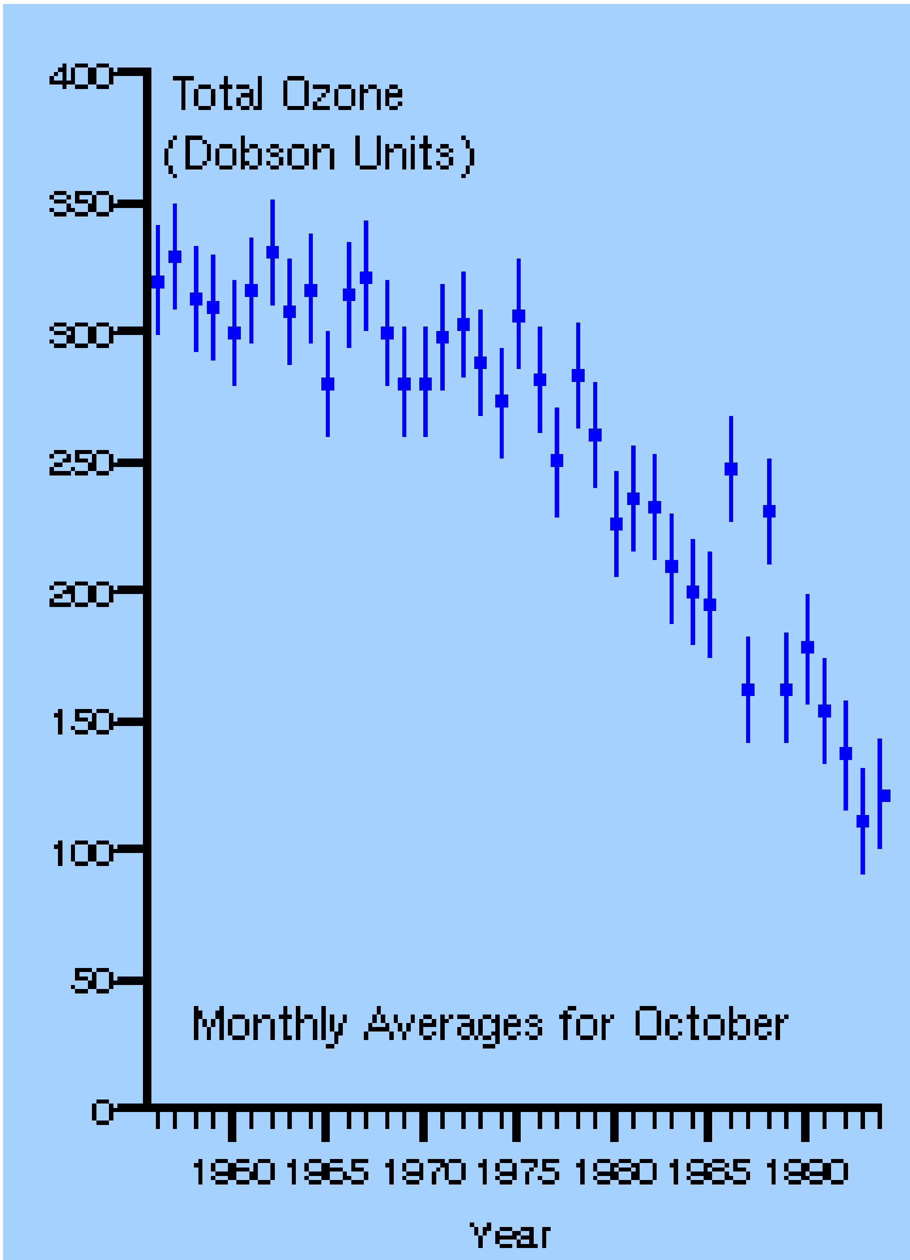

Figure 1.16: Monthly average of ozone concentration in October over Antarctica. From www.atm.ch.cam.ac.uk/tour/

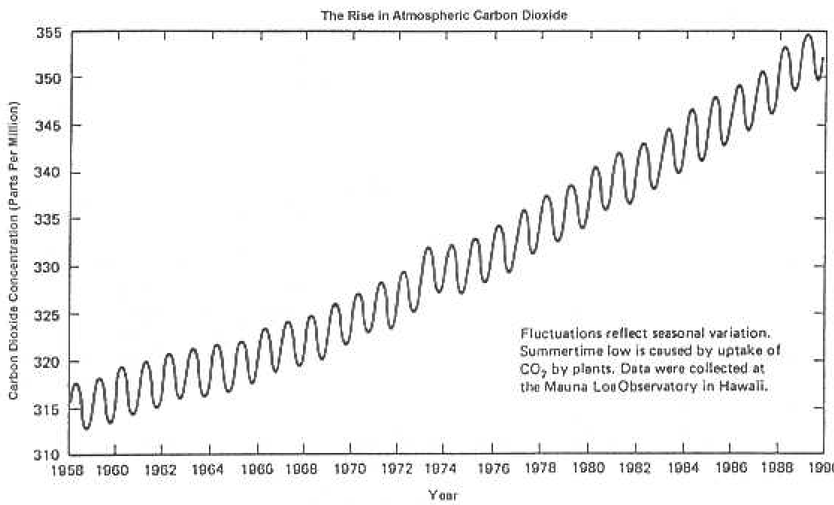

Figure 1.17: CO_2 concentration measured in the Pacific.

$$S = 1.376\text{kWm}^{-2} \tag{1.8}$$

Thus a 1 m diameter dish in space would collect about 1 kW; enough to run a small electric heater.

The total energy received from the sun per unit time is

$$E = \pi R^2 S \tag{1.9}$$

where R is the radius of the earth. Since the surface area of the earth is $4\pi R^2$, the average amount of energy received per unit area per unit time at the earth's surface is

$$\frac{1}{4} S = 344\text{Wm}^{-2} \tag{1.10}$$

If the earth's axis were not tilted the average flux would vary from $\pi^{-1} S$ at the equator to zero at the poles. However, since the axis is tilted (23.5^o) the average flux is as shown in figure 1.19.

Not all the incoming energy is absorbed. A fraction $\overline{\alpha}$ is reflected or scattered, and so the average flux absorbed is

$$\frac{1}{4}\left(1 - \overline{\alpha}\right) S = 240\text{Wm}^{-2} \tag{1.11}$$

The amount reflected or scattered is about 100 Wm^{-2}, and is fairly independent

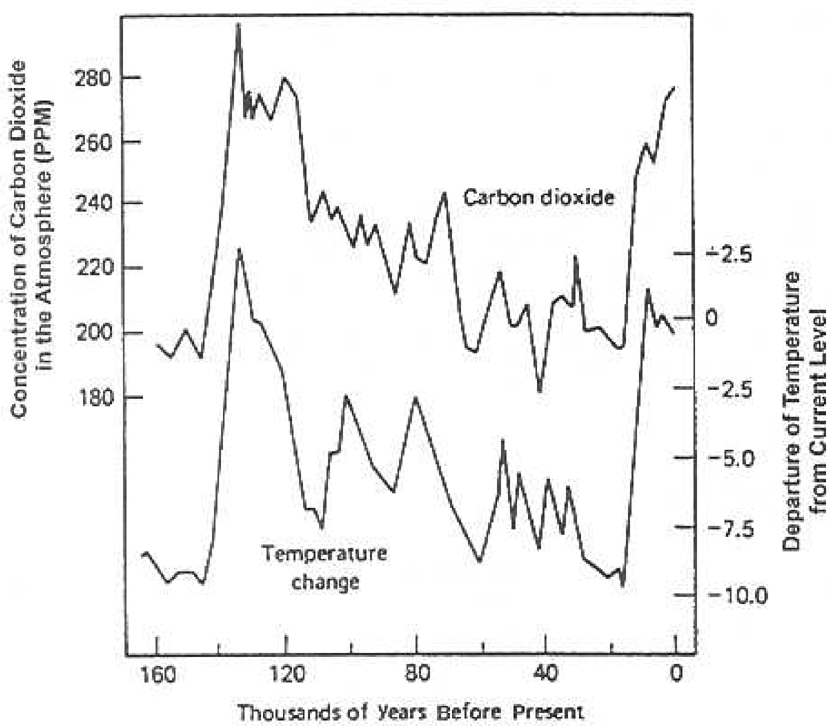

Figure 1.18: CO_2 concentration and temperature change from the present over the past 160,000 years.

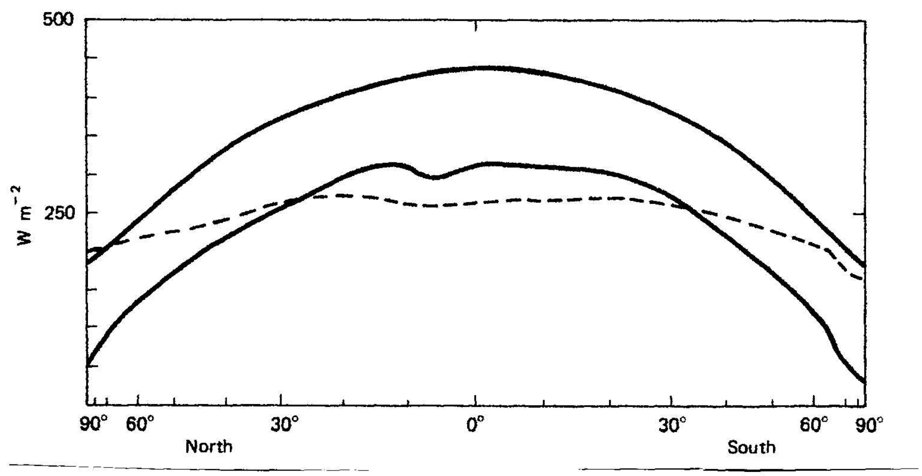

Figure 1.19: Incoming radiation as a function of latitude. The upper solid curve is the average incoming radiation and the lower solid curve is the average amount absorbed. The dashed curve is the amount of outgoing radiation.

of latitude. The number $\overline{\alpha}$ is called the *albedo* [3]of the earth and has a value of about $\overline{\alpha} = 0.3$.

1.3.2 Radiative equilibrium models

The absorption of energy causes the surface to warm up until it radiates the same amount of energy to space. When the surface reaches temperature T, the amount of energy E radiated is given by Stefan's law [4]

[3]Albedo is the "fraction of light that is reflected by a body or surface. It is commonly used in astronomy to describe the reflective properties of planets, satellites, and asteroids. Albedo is usually differentiated into two general types: normal albedo and bond albedo. The former, also called normal reflectance, is a measure of a surface's relative brightness when illuminated and observed vertically. The normal albedo of snow, for example, is nearly 1.0, whereas that of charcoal is about 0.04. Investigators frequently rely on observations of normal albedo to determine the surface compositions of satellites and asteroids. The albedo, diameter, and distance of such objects together determine their brightness. Bond albedo, defined as the fraction of the total incident solar radiation reflected by a planet back to space, is a measure of the planet's energy balance. (It is so named for the American astronomer George P. Bond, who in 1861 published a comparison of the brightness of the Sun, the Moon, and Jupiter.) The value of bond albedo is dependent on the spectrum of the incident radiation because such albedo is defined over the entire range of wavelengths. Earth-orbiting satellites have been used to measure the Earth's bond albedo. The most recent values obtained are approximately 0.33. The Moon, which has a very tenuous atmosphere and no clouds, has an albedo of 0.12. By contrast, that of Venus, which is covered by dense clouds, is 0.76." (from a google search leading to a web site at the University of Oregon). Albedo features of Mars may be viewed at http://mars.jpl.nasa.gov/MPF/mpf/marswatch/marsnom.html

[4]The law (and constant) were discovered in 1879 by the Austrian physicist Josef Stefan (183593)

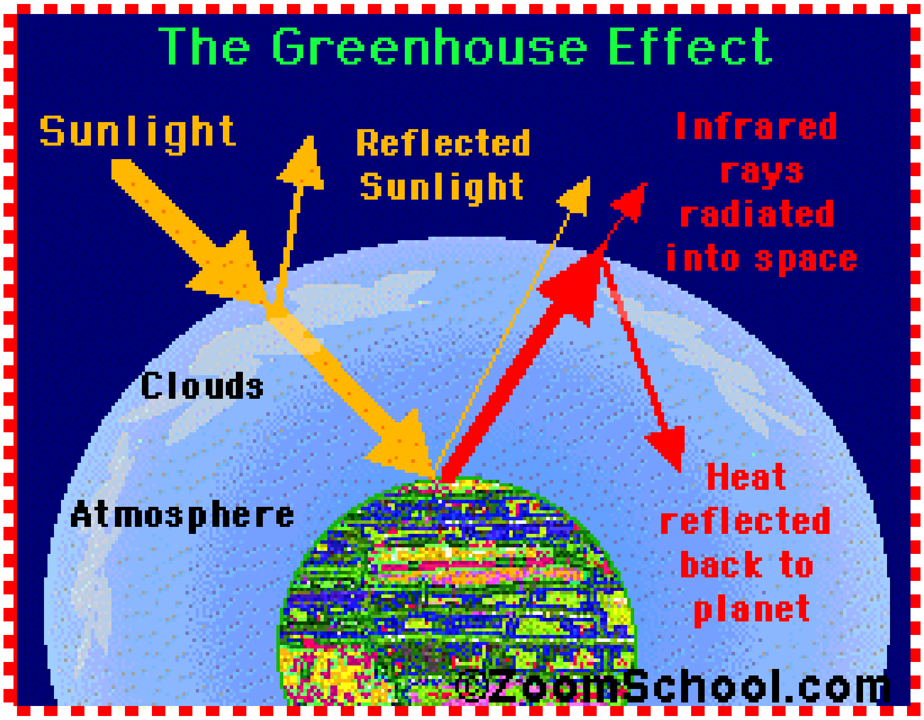

Figure 1.20: Schematic of the greenhouse effect. From www.enchantedlearning.com

$$E = \sigma T^4, \tag{1.12}$$

where $\sigma = 5.7 \times 10^8$ Wm^{-2} K^{-4} . Such an equilibrium would give a temperature at the equator of 270 K, and 150 K at the South Pole and 170 K at the North Pole. The difference between these temperatures and those on the earth are a result of the atmosphere and oceans:

- (a) some radiation can be absorbed within the atmosphere itself
- (b) the atmosphere and oceans can transport heat from one location to another.

1.3.3 The greenhouse effect

A schematic of the radiation into and from the earth's atmosphere is shown in figure 1.20. Infrared radiation is trapped in the atmosphere by greenhouse gases, mainly CO_2, methane and water vapour. Consider a 'greenhouse' modelled as a sheet of glass above the ground as shown in figure 1.21. The glass is

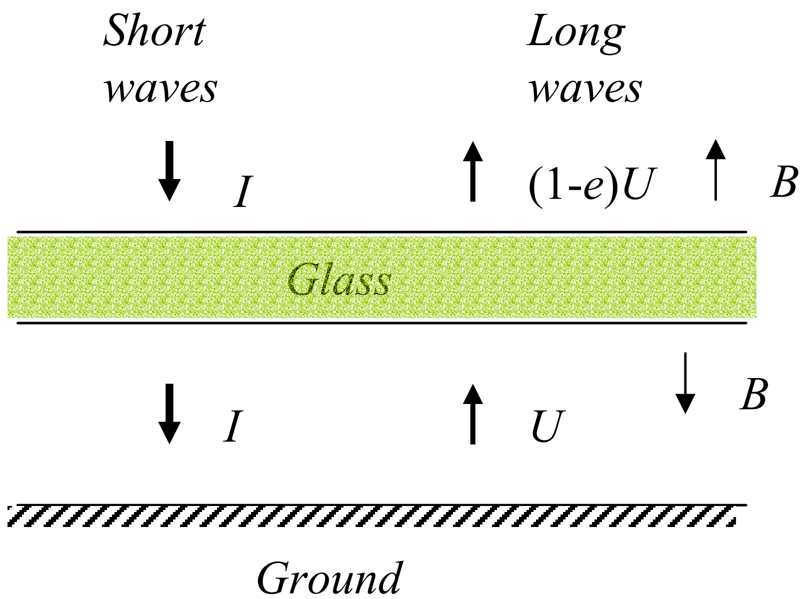

Figure 1.21: Schematic of the radiation balance caused by greenhouse gases.

transparent to wavelengths less than 4mm, and partially absorbs radiation of longer wavelengths. Initially the glass is cold, and a downward flux I of solar radiation is switched on.

The ground will warm up to a temperature T_g, and emit long wave radiation with an upward flux U given by Stefan's law [5]

$$U = \sigma {T_g}^4. \tag{1.13}$$

Almost all of the radiation has wavelengths above 4 mm (the range is 4-100 mm at typical temperatures), and a fraction e of this will be absorbed by the glass. The glass will also warm up and emit radiation B.

Equilibrium is reached when the upward and downward fluxes match

$$I = (1 - e)U + B = U - B. \tag{1.14}$$

Solving (1.13) and (1.14) gives

[5]Joseph Stefan (Slovene Joef Stefan) (1835 – 1893) was a Slovene mathematician, physicist and poet. *'There always something will remain, that we shall not know, why?'*

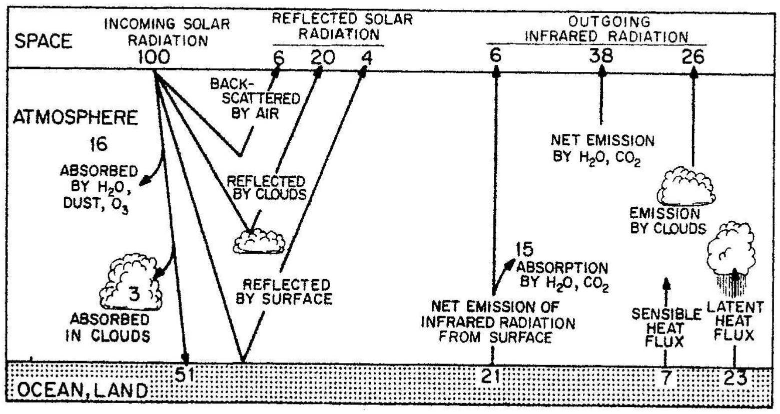

Figure 1.22: Radiation balance of the atmosphere.

$$\sigma {T_g}^4 = U = I(1 - \frac{e}{2})^{-1}. \quad (1.15)$$

Thus T_g is higher than it would be in the absence of the glass ($e = 0$).

Consider the case where the glass absorbs all the long wave radiation ($e = 1$). Then $I = B$, and the glass would reach the same temperature as the ground in the absence of glass. Since the underside of the glass is at the same temperature, it radiates a downward flux B of long wave radiation, so the ground receives a total flux of $I + B = 2I$. Thus by Stefan's law the temperature is higher by $2^{1/4} = 1.19$. In the atmosphere the absorption occurs over a range of heights but the principle is the same (Chamberlain (1978)). An estimate of the radiation balance is given in figure 1.22.

Problem 1.1 *Figure 1.15 is a schematic of the development of the density profile near the ground corresponding (approximately) to the temperature profiles shown in figure 1.14. (a) is the initial profile. Calculate the change in potential energy per unit area of the stratification in cases (b) and (c) in terms of the initial density gradient ρ_z and the height h. Assuming that the difference between (b) and (c) is due to solar radiation, explain why the change in potential energy is positive in one case and negative in the other.*

Problem 1.2 *Find the value of c_p for air and for water. Hence calculate the adiabatic lapse rate for both the atmosphere and the ocean. Given that the temperature gradient in the lowest 11km of the atmosphere is -6.5 K km^{-1}, find the effective temperature gradient – called the* potential temperature gradient *– once the effects of compressibility have been removed.*

Problem 1.3 *Stefan's law (1.12) is for a 'black body' (i.e. a perfect emitter). A more*

general form is $E = e_m \sigma T^4$, *where* $0 \leq e_m \leq 1$ *is the* emissivity *of the emitting surface. Calculate the greenhouse enhancement of the surface temperature for*
(a) $e_m = 0.5, e = 1$;
(b) $e_m = 1, e = 0.5$,
where e *is the emissivity of the atmosphere. Comment on the implications for the earth's climate of changing the characteristics of the earth surface and the emissivity of the atmosphere.*