MBARARA UNIVERSITY OF SCIENCE AND TECHNOLOGY

FACULTY OF SCIENCE

DEPARTMENT OF PHYSICS

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Chapter 1

Introduction

The Earth's natural processes operate across various scales, from local phenomena like earthquakes to continent-scale motions over thousands to millions of years. Understanding how the Earth interacts with air, water, and life is crucial for grasping its dynamics. Over its 4.6 billion years, Earth has developed a complex environment shaped by interactions between land, air, oceans, and life. However, humanity's pursuit of prosperity has led to exploitation of natural resources, resulting in environmental issues like ozone depletion, global warming, acid rain, and urban pollution.

Addressing these challenges requires interdisciplinary approaches, as environmental problems transcend traditional academic disciplines. Environmental physics is an interdisciplinary subject that integrates the physics processes in the following disciplines: the atmosphere, the biosphere, the hydrosphere, and the geosphere. It examines the response of living organisms to their environment within the framework of environmental processes and issues. Environmental physics can be defined as the response of living organisms to their environment within the framework of the physics of environmental processes and issues. It is structured within the relationship between the atmosphere, the oceans (hydrosphere), land (lithosphere), soils and vegetation (biosphere). It embraces themes such as human environment and survival physics, built environment, urban environment, renewable energy, remote sensing, weather, climate and climate change, and environmental health. This interdisciplinary approach allows for a comprehensive understanding and assessment of environmental challenges and their solutions.

In subsequent chapters, the application of physics principles to environmental processes and problems will be explored and contextualized.

Chapter 2

The human environment

Living organisms, including humans, must adapt to diverse environmental conditions, ranging from extreme cold to intense heat. Their ability to regulate body temperature within a narrow range, typically 35-40°C, is crucial for survival. The physics governing energy transfers and thermoregulation include fundamental principles such as the laws of thermodynamics, concepts of entropy, enthalpy, and Gibbs free energy, as well as mechanisms like conduction, convection, radiation, and evaporation. Additionally, Newton's law of cooling and Wien's and Stefan-Boltzmann radiation laws play essential roles in understanding these processes.

Humans and other mammals possess the remarkable ability to maintain a constant body temperature, termed homeothermy, despite varying environmental conditions. They achieve this by adjusting energy transfer and production rates. In contrast, certain animal species like reptiles and amphibians, known as poikilotherms, have body temperatures that fluctuate with environmental changes.

Both homeotherms and poikilotherms employ physiological and behavioral mechanisms to adapt to their surroundings. Humans, for instance, wear warmer clothing in cold weather and lighter clothing in hot climates. Understanding the physics of thermal processes is crucial for sustaining life on Earth, as it complements knowledge of metabolic reactions in chemistry and biochemistry. Exploring the laws of thermodynamics helps elucidate how these principles apply to the body's energy metabolism and overall survival strategies.

2.1 Laws of thermodynamics

2.1.1 First law of thermodynamics

The First Law of Thermodynamics, a fundamental principle in physics, states that energy cannot be created or destroyed in an isolated system. Instead, it can only change forms or be transferred from one part of the system to another, or between the system and its surroundings. Mathematically, the first law is often expressed as:

$$\triangle \mathrm{U} = \triangle \mathrm{Q} - \triangle \mathrm{W} \tag{2.1}$$

where ($\triangle$U) represents the change in internal energy of the system, ($\triangle$Q) is the heat added to the system, and ($\triangle$W) is the work done by the system. In other words, the change in internal energy of a system is equal to the heat added to the system minus the work done by the system.

This law is a statement of the principle of energy conservation, which asserts that the total energy of an isolated system remains constant over time. It provides a foundation for understanding the flow and transformation of energy in various physical and chemical processes, from the behavior of gases to the operation of heat engines.

Enthalpy, denoted as $\triangle$H, is another essential concept in thermodynamics. It represents the heat content of a system and is a thermodynamic state function. Enthalpy is related to the internal energy ($\triangle$U), pressure p, and volume V by the equation

$$\mathrm{H} = \mathrm{U} + \mathrm{pV} \tag{2.2}$$

Often, it's more practical to discuss the enthalpy change $\triangle$H of a chemical reaction. When no external work is accomplished ($\triangle \mathrm{W} = 0$), $\triangle \mathrm{H} = \triangle \mathrm{U}$. This enthalpy change quantifies the amount of energy generated or absorbed during a reaction.

2.1.2 Second law of thermodynamics

Both an internal combustion engine and the human body operate as heat engines, extracting useful mechanical work from systems with temperature differences between their interiors and surroundings. This analogy helps understand how our bodies function. The operation of any heat engine is governed by the Second Law of Thermodynamics, originally formulated by the French physicist Sadi Carnot.

According to Carnot's principle, in a heat engine, work is obtained from the energy transferred between a higher temperature body and a lower temperature one. This energy transfer cannot occur spontaneously in the opposite direction without external intervention. The Second Law is often expressed in terms of efficiency, given by the formula

$$\mathrm{e} = \frac{\mathrm{T}_1 - \mathrm{T}_2}{\mathrm{T}_1}, \tag{2.3}$$

where T_1 represents the higher temperature and T_2 the lower temperature.

The significance of the Second Law lies in its definition of the direction in which thermal energy flows. It clarifies that heat naturally moves from regions of higher temperature to regions of lower temperature, elucidating the constraints and possibilities within thermal systems.

2.1.3 Third law of thermodynamics

The cooling of a cup of tea left in a room is a simple example illustrating the Second Law of Thermodynamics. The tea, initially at 60°C, gradually cools to match the lower temperature of its surroundings, 20°C. This process is irreversible without external intervention, highlighting the natural flow of heat from higher to lower temperatures. Similarly, the human body relies on external sources like food for chemical energy and solar radiation to maintain its temperature. Without these inputs, the body's temperature would decrease, leading to eventual death.

The temperature difference between our bodies and the environment not only sustains us but also allows for the production of useful mechanical work. As energy flows out of the body into the environment, the process is irreversible, and the environment gains energy $\triangle \mathrm{Q}$ at its temperature $\triangle \mathrm{T}$. This exchange defines the entropy change $\triangle \mathrm{S}$, where

$$\triangle \mathrm{S} = \frac{\triangle \mathrm{Q}}{\mathrm{T}}. \tag{2.4}$$

The combined entropy change of the body and environment is always greater than zero, emphasizing the increase in disorder associated with natural processes.

Entropy, defined by Ludwig Boltzmann, relates to the probability W of energy distributions within a system. Boltzmann's equation

$$\mathrm{S} = \mathrm{k} \cdot \ln \mathrm{W}, \tag{2.5}$$

where k is Boltzmann's constant, elucidates the relationship between entropy and energy distribution probabilities.

The third law of thermodynamics, which posits that entropy is zero at absolute zero temperature (0 K), allows for the determination of entropy changes. For example, the entropy change when ice at 0°C transitions to liquid water at 0°C can be calculated using $\triangle \mathrm{S} = \frac{\mathrm{dQ}}{\mathrm{T}}$. Entropy serves as a measure of the 'disorder' within a system, with most natural processes leading to an increase in entropy. Irreversible processes, such as the cooling of tea or the decay of radioactive sources, contribute to the overall increase in entropy within the system.

2.2 Laws of thermodynamics and the human body

The Second law governs changes that act in the direction in which entropy increases. We will now see through a detailed examination how the laws of thermodynamics relate to the energetics of the body.

2.2.1 Energy and metabolism

Metabolism encompasses all chemical processes occurring within the cells of an organism, comprising anabolism, where molecules are synthesized, and catabolism, where enzymes break down food molecules through hydrolysis. At the cellular level, metabolism involves phosphorolysis. The *basal metabolic rate* (BMR) denotes the rate at which a fasting, sedentary body generates enough energy to sustain vital functions like respiration, temperature regulation, heart rate, and tissue production. BMR, is approximately equal to the metabolic rate during sleep, and while resting most of the energy is dissipated as thermal energy.

BMR can be determined using direct calorimetry or spirometry, where energy production correlates with oxygen consumption during respiration. For instance, a man's

BMR is approximately 170 kJ/m^2/h, while a woman's is about 155 kJ/m^2/h. Considering typical daily activities, energy dissipations vary from sleeping (75 W) to running hard (400-1500 W). An average person may require around 4200 kJ for a typical working day, totaling approximately 12000 kJ/day. Adjusting macronutrient intake (carbohydrates, proteins, fats) can meet these energy needs.

Metabolism involves chemical processes transferring energy between compounds, generating thermal energy. Increased metabolic reaction rates correspond to increased energy generation. Energy requirements vary based on tasks, influencing athletic performance and survival. A sedentary man typically produces around 0.07 kJ/kg/min (equivalent to about 80 W for a 70 kg man). Understanding metabolism sheds light on energy utilization, performance, and health maintenance in individuals.

2.2.2 Thermodynamics and the human body

Humans breathe in oxygen and eat food, which is composed of carbohydrates, fats, oils and proteins. The carbohydrates are converted into glucose, the proteins into amino acids, and the fats into fatty acids. The blood then transports these, together with oxygen, to the cells, where enzymes, which are biological catalysts, convert the glucose into pyruvic acid, through the process of glycolysis. The fatty and most of the amino acids are converted into acetoacetic acid. These are changed into acetyl Co-A, and with further ocidation, produce adenosine triphosphate (ATP), carbon dioxide and water. This entire process is called the Krebs Cycle.

ATP serves as the primary energy currency for cells, with energy stored in phosphate bonds during the conversion of ADP to ATP and released when ATP reverts to ADP. This released energy primarily dissipates as heat, transferred throughout the body via blood circulation. Thermal energy loss from the body occurs through conduction, convection, radiation, and evaporation from the skin, as well as through respiration.

Humans maintain a stable core body temperature within a narrow range of 35-40°C, with the normal core temperature being 37°C. However, temperature gradients exist both within the body and between the body and its surroundings, leading to heat exchange processes to regulate body temperature.

Physics is crucial in understanding energy and metabolism because it underpins the biochemical processes that provide energy to the body. While this discussion doesn't delve deeply into biochemical processes, physics, especially through the laws of thermodynamics, plays a significant role in metabolic processes. Thermodynamic principles govern energy transfer, transformation, and dissipation within biological systems, providing a foundational understanding of energy metabolism and its regulation. Thus, physics offers essential insights into the energetic dynamics of living organisms, including humans.

2.2.3 First law of thermodynamics and the human body

In an energy balance scenario, where the core body temperature and ambient temperature remain constant, the energy produced by the body equals the energy dissipated. This equilibrium allows for the application of the First Law of Thermodynamics to the body. The total energy produced within the body, known as the metabolic rate ($\triangle$M), is related to both the total metabolic energy production ($\triangle$H) and the external work done by the body ($\triangle$W).

The relationship is expressed as:

$$\triangle \mathrm{M} = \triangle \mathrm{H} + \triangle \mathrm{W}. \tag{2.6}$$

This equation bears analogy to the expression of the First Law of Thermodynamics. ($\triangle$H) varies among individuals and depends on the activity level and body surface area. On average, the body's surface area is about 1.84 m^2, with the average male mass around 65-70 kg and the average female mass around 55 kg. A sedentary person typically has a metabolic rate of about 100 W, while someone engaged in heavy physical work may have a metabolic rate of 400 W.

Energy transfers within metabolic processes adhere to the First Law of Thermodynamics. If no mechanical work is performed ($\triangle \mathrm{W} = 0$), then the chemical energy input is transferred as thermal energy. In other words, $\triangle \mathrm{H} = \triangle \mathrm{U}$, where $\triangle$U represents the energy produced by the oxidation of chemicals, and the total $\triangle$H is the product of the mass of chemicals oxidized and $\triangle$H.

This application of the First Law helps quantify the energy generated within the body's metabolic processes, offering insights into energy utilization and thermodynamic principles governing physiological functions.

2.2.4 Second law of thermodynamics and the human body

If a metabolic process occurs in a particular direction, does it also occur in the reverse manner? The Second law helps to explain both the direction and attainment of equilibrium in metabolic processes, and now it can be seen that the entropy change can assist in the understanding of the direction that a metabolic process will take. It also tells us whether that particular process will occur.

In the oxidation of glucose, some energy is "wasted" as heat, rendering the process less than 100% efficient. This "wasted" energy, crucial for maintaining core body temperature, serves as the driving force behind the direction of metabolic processes. The concept of potential energy aids in understanding the direction of these processes. For instance, when an object falls, its potential energy converts into kinetic energy and eventually dissipates as heat, sound, and possibly light, leading to an increase in entropy of the surroundings (i.e., the Universe). The change in entopy $\triangle S_{environment}$ due to energy transfer from the body is expressed as

$$\triangle \mathrm{S}_{\mathrm{environment}} = -\frac{\triangle \mathrm{Q}_{\mathrm{body}}}{\mathrm{T}} \tag{2.7}$$

under the assumption of isothermal conditions at the cellular level. A decrease in energy from the body $\mathrm{Q}_{\mathrm{body}}$ corresponds to an increase in $\triangle \mathrm{S}_{\mathrm{environment}}$.

Entropy serves as a guide for the direction of spontaneous changes. Establishing criteria for the propensity of a system to provide "free energy" for useful work is essential. The concept of Gibbs free energy (GG) offers this criterion. Gibbs free energy represents the maximum possibility of a process achieving work. If dGdG is negative, free energy is released, enabling the occurrence of the process. Conversely, if dGdG is positive, the process will not occur.

Since the First law of thermodynamics can be represented as

$$\triangle \mathrm{Q} = \triangle \mathrm{U} + \mathrm{p} \cdot \triangle \mathrm{V}, \tag{2.8}$$

where p and $\triangle$V is the pressure and the change in volume, and the Second law by

$$\triangle \mathrm{S} = \frac{\triangle \mathrm{Q}}{\mathrm{T}} \tag{2.9}$$

then

$$\mathrm{T} \cdot \triangle \mathrm{S} = \triangle \mathrm{U} + \mathrm{p} \cdot \triangle \mathrm{V}, \tag{2.10}$$

where $\triangle S$ is the change in entropy related to a change in energy, $\triangle Q$. Therefore, the change in internal energy is

$$\triangle \mathrm{U} = \mathrm{T} \cdot \triangle \mathrm{S} - \mathrm{p} \cdot \triangle \mathrm{V}. \tag{2.11}$$

This is the Gibbs equation. It incorporates the idea of temperature, it embraces the Zeroth, the First and Second laws of thermodynamics. The temperature is a central characteristic of a thermodynamic system. It can be applied to any physical systems (and to the biophysical system that is the human body). Using the definition of enthaly, H = U + pV, we have

$$\triangle \mathrm{H} = \triangle \mathrm{U} + \mathrm{p} \cdot \triangle \mathrm{V} + \mathrm{V} \cdot \triangle \mathrm{p} = \mathrm{T} \cdot \triangle \mathrm{S} + \mathrm{V} \cdot \triangle \mathrm{p}. \tag{2.12}$$

Now, we define the Gibbs free energy, G, as

$$\mathrm{G} = \mathrm{U} - \mathrm{TS} + \mathrm{pV}. \tag{2.13}$$

Thus, it is easy to see that the change in the Gibbs free energy is

$$\triangle \mathrm{G} = \triangle \mathrm{H} - \mathrm{T} \cdot \triangle \mathrm{S}. \tag{2.14}$$

The above equation gives the maximum possibility of a process achieving work. G is not free energy in the sense that it comes from nothing. It implies that it is the energy available for work. $\triangle G$ influences the possible direction of a metabolic process. If it is negative, then free energy is released and the process will occur. If it is positive, it will not.

2.3 Energy transfers

To feel warm, whether one is in the house or walking outside, is a question of energy conservation, but the underlying principle is that of an energy balance, and for this to be achieved energy exchange is necessary. Energy can be transferred from one point to another by the following mechanisms: conduction, convection, radiation, and evaporation. The physics of each of those mechanisms will be discussed in turn.

2.3.1 Conduction

Thermal conduction is the process by which energy transfers between two points in a material at different temperatures. In solids, this occurs through two mechanisms: (i) molecular vibrations transmitting energy through the crystal lattice and (ii) the mobility of free conduction electrons within the lattice. In semiconductors, both mechanisms contribute, while in insulators, the first mechanism predominates. These lattice vibrations, known as phonons, create elastic acoustic standing waves that propagate through the material at the speed of sound for that material.

Joseph Fourier discovered that the rate of thermal energy flow (dQ/dt) through a material depends on its cross-sectional area (A), length or thickness (L), and the temperature difference between the two sides ($\triangle T = T_1 - T_2$). This relationship can be expressed as:

$$\frac{\mathrm{dQ}}{\mathrm{dt}} = -\mathrm{kA} \cdot \frac{\triangle \mathrm{T}}{\mathrm{L}}, \tag{2.15}$$

where k is the thermal conductivity of the material. Thermal conductivity determines a material's effectiveness as an insulator or conductor. Materials like copper, with high thermal conductivity (e.g., 380 $Wm^{-1}K^{-1}$), conduct heat well, while poor conductors like water have low thermal conductivity (e.g., 0.59 $Wm^{-1}K^{-1}$). The ratio of temperature difference to length is termed the temperature gradient.

The negative sign in the equation indicates that energy flows from the region at the higher temperature to the lower temperature, following the temperature gradient, making energy flow unidirectional. This equation holds true for steady-state conditions, where temperatures are stable, and thermal energy input equals thermal output, or for short time intervals.

2.3.1.1 Example: Fourier's law of thermal conduction.

A rambler is walking up a steep hillside in January. He is wearing clothing 1 cm thick, his skin temperature is 34^0C and the exterior surface is close to freezing at 0^0C. Determine the rate of flow of energy outwards from his body, through thermal conduction, when:

(i) it is fine dry. Assume that the thermal conductivity for clothing, under dry conditions, is 0.042 $Wm^{-1}K^{-1}$.

(ii) It has been raining heavily and the rambler is soaked. The thermal conductivity is now 0.64 $Wm^{-1}K^{-1}$. Assume that the walker has surface area of 1.84 m^2.

Solution:

(i) Apply the Fourier's law of thermal conduction.

dQ/dt = - $k \cdot A \cdot dT/L = -0.042 \cdot 1.84 \cdot 34/0.01$ W = - 263 W.

(ii) Applying the Fourier's law again, dQ/dt = - 4004 W.

When clothing becomes wet, it becomes a better conductor for the outward dissipation of energy because water has a higher thermal conductivity than dry clothing. This is why jeans are inappropriate trouser-ware for strenuous outdoor pursuits in wet weather.

2.3.2 Convection

Convection is the process through which thermal energy is transferred by the movement of a fluid, which can be either a liquid or a gas. When a living body radiates heat, the air around it warms up, becomes less dense, and rises, while colder, denser air moves downward, creating a convection current. This process also occurs on a larger scale in the Earth's atmosphere. There are two types of convection: natural (occurring without external forcing) and forced (induced by external factors, such as blowing air over a hot surface). Forced convection, particularly relevant in human environments, is described by Newton's law of cooling.

2.3.2.1 Newton's Law of Cooling

Newton's law of cooling states that the rate of energy loss (dQ/dt) from an object in a fluid is directly proportional to the temperature difference (△T) between the object and the surrounding environment. Mathematically, it can be expressed as:

$$\frac{\mathrm{dQ}}{\mathrm{dt}} = -\mathrm{kA} \cdot \triangle \mathrm{T} \tag{2.16}$$

Where:
- k is the convective energy transfer coefficient, depending on the nature and surface

area of the object.
- A is the surface area of the object.
- $\triangle$T is the temperature difference between the object and the surrounding environment.

For example, for a plate in still air, k is approximately 4.5 $Wm^{-1}K^{-1}$, while with air flowing over it at 2 m/s, k increases to about 12 $Wm^{-1}K^{-1}$. While Newton's law of cooling doesn't directly apply to humans, as their metabolism attempts to maintain a constant body temperature, walking against strong winds or in a wind tunnel approximates this law well.

2.3.2.2 Example: Fluid flow and convection.

A student volunteers to take part in the following simulations of convective energy loss:

(a) he is placed in a wind tunnel in which air, at -20C, blows through at 40 km/h.
(b) He is placed in a flow of water, which is at 120C. The velocity of flow is 0.5 m/s.
In each case calculate:

(i) the convective energy transfer coefficient, and

(ii) the convective energy transfer flux, dQ/dt. Assume A = 1.8 m^2 and the skin temperature is 31^0C. Assume convective energy transfer coefficient k = 44.8 $Wm^{-1}K^{-1}$ for 40 km/h speed of flow, and k = 34.4 $Wm^{-1}K^{-1}$ for 0.5 m/s.

Solution:

Applying Newton's law for cooling, we have:
(a) dQ/dt = 2661 W.
(b) dQ/dt = 1176 W.

2.3.3 Radiation

Radiation plays an important role in the energy balance of human beings. It is the process in which energy can be transferred in the form of electromagnetic waves from one point to another through a vacuum. All objects release energy in the form of electromagnetic waves. The best absorbers usually make the best emitters of radiation and these are called black-body. Human beings emit radiation in the infrared band. There are two experimental laws that usually we use to explain radiation: Wien's and Stefan-Boltzmann laws. Wien's law tells us about the wavelength, λ_{m}, that the body at the temperature, T, radiates with the maximum intensity

$$\lambda_{\mathrm{m}} \cdot \mathrm{T} = \mathrm{b}, \tag{2.17}$$

where b is a constant, and for the black-body is $\mathrm{b} = 3 \cdot 10^{-3} m \cdot K$. Stefan-Boltzmann law explains the total amount of radiation energy per second (or as power) from a black-body, and was discovered to be proportional to the forth power of the temperature, T, and the area, A, of the surface emitting the radiation

$$\mathrm{P} = \sigma \mathrm{A} \mathrm{T}^4, \tag{2.18}$$

where σ is Stefan's constant, and for the black-body is $\sigma = 5.7 \cdot 10^{-8} \mathrm{Wm}^{-2}\mathrm{K}^{-4}$. No real body is a perfect black-body radiator, and to distinguish between a perfect black-body and real bodies the idea of emissivity, ε, has been introduced. Thus, for the radiation power we have:

$$\mathrm{P} = \varepsilon \sigma \mathrm{A} \mathrm{T}^4 \tag{2.19}$$

An object release energy in an environment which itself is radiating energy. Then the object will absorb energy. This means that the net rate of radiation emitted is the difference between the energy emitted by the object and the energy absorbs from its surroundings.

2.3.3.1 Example: Radiation and the Stefan-Boltzmann law.

A person sitting reading a book releases radiant energy of between 70 and 100 W. Calculate how much energy the person is radiating. Assuming that the emissivity for the human body is 0.5, with an average surface temperature of 35^0C, that room temperature is 20^0C, and the body surface area is 1.8 m^2.

Solution:
Applying the Stefan-Boltzmann law, P = 83.6 W.

2.3.4 Evaporation

Evaporation plays a crucial role in the process of maintaining the body's energy balance, particularly during strenuous physical activities when metabolic rates increase. This process allows the body to dissipate excess heat and maintain a stable core temperature. Here's a breakdown of the concept of evaporation and its significance in human thermoregulation:

Evaporation is the process by which a liquid transforms into a vapor. It involves a phase change and is characterized by a latent heat change, where energy is required to vaporize the liquid.

$$\mathrm{Q} = \mathrm{mL}, \tag{2.20}$$

where Q is the energy extracted or supplied to bring about a phase change, m is the mass of liquid to be vaporized and L is the specific latent heat of vaporization. For pure water, $\mathrm{L} = 2.25 \cdot 10^6$ J/kg, but sweat, which is 99% water with sodium chloride as solute, is an electrolyte with $\mathrm{L} = 2.43 \cdot 10^6$ J/kg. For humans and animals L depends on temperature.

The energy required for evaporation is obtained from the warmer body. As sweat evaporates from the skin's surface, it absorbs heat energy, resulting in a cooling effect on the body. This cooling effect helps regulate body temperature, especially during physical exertion.

The rate of evaporation depends on several factors:
- Surface Area: Larger surface areas allow for more efficient evaporation.
- Temperature Difference: Higher temperature differences between the body and its surroundings facilitate faster evaporation.
- Humidity: Lower humidity levels create a greater difference in vapor pressure, promoting faster evaporation.
- Sweating Rate: The rate at which sweat is produced influences the amount of moisture available for evaporation.
- Air Velocity: Faster air movement enhances the rate of evaporation by removing vapor-saturated air from the skin's surface.

The rate of evaporation (dQ/dt) can be expressed as:

$$\frac{\mathrm{dQ}}{\mathrm{dt}} = \mathrm{hA}(\mathrm{p_S} - \mathrm{p_0}) \tag{2.21}$$

where:
- h is the evaporative energy transfer coefficient,
- A is the surface area of the skin,
- p_S is the water vapor pressure adjacent to the skin, and
- p_0 is the water vapor pressure in the surrounding air.

Unlike non-evaporative energy loss mechanisms such as conduction and convection, which rely on temperature gradients, evaporation relies on vapor pressure gradients. In environments with high humidity, evaporative cooling may be less efficient due to reduced vapor pressure gradients.

During rest, a thermally comfortable individual may experience evaporative water loss of around 30 g/h. However, during intense physical activity, sweating rates can increase significantly, resulting in higher rates of evaporative water loss (e.g., up to 200 g/h).

2.3.5 Survival in cold climates

As individuals age, their metabolic rate decreases, leading to a reduced capacity to generate heat. In cold weather, especially without adequate insulation and heating, older individuals may be at risk of hypothermia due to the imbalance between heat production and dissipation. Conversely, during very hot summers, babies and the elderly are susceptible to heat stress due to challenges in thermoregulation.

Wind, like water, behaves as a fluid and can influence the transfer of thermal energy from the body. Wind flow can vary between laminar and turbulent, affecting convective airflow patterns. Wind-chill temperature, determined by wind speed and air temperature, can decrease as wind speed increases. Exposure to wind, especially in open areas like hills, can enhance heat loss from the body, potentially leading to discomfort and headaches.

Hypothermia occurs when the body's core temperature falls below normal levels, presenting a shift from negative to positive biological-physical feedback mechanisms. Initially, the body compensates for cold temperatures through increased energy production. However, if the temperature continues to drop, the body's ability to generate heat becomes insufficient, leading to progressive hypothermia. Hypothermia progresses through stages, starting with mild symptoms and potentially leading to death in severe cases. Immersion in cold water, particularly in windy conditions, can accelerate heat loss and increase susceptibility to hypothermia, especially in lean individuals.

2.3.6 Survival in hot climates

If energy transfers from the environment into the human body, without some dissipating mechanism the body's temperature will increase, to the point of heat stress, and beyond to heat stroke and death. The balance can be brought to an equilibrium steady-state, by perspiration in humans and panting in some animals. Energy from the body is used to vaporize sweat and a cooling effect results. Working in very hot climates, such as deserts, can results in a water loss of 10-12 kg/day. In hot climates evaporative cooling becomes the dominant mechanism of energy transfer and thermoregulation.

Like hypothermia, heat stress or hyperthermia has stages, all of which are accompanied by continuing dehydration:
sweating and vasodilatation: Muscular activity can disrupt the body's energy balance, leading to increased body temperature. Vasodilation (widening of blood vessels) and sweating are the body's initial responses to dissipate heat.
Heat cramp: Continued dehydration can lead to heat cramps, affecting muscles, particularly in the legs and stomach. Dizziness may also occur.
Heat exhaustion: Heat exhaustion sets in when water and salt are not adequately replenished after sweating. Symptoms include dehydration and further elevation of core body temperature.
Heat stroke: This is very serious and can occur when the body's temperature is in excess of 41^0C. The thermoregulatory system, especially the process of vasodilation, starts to collapse, with the result that the body cannot effectively dissipate energy and, therefore, the body's temperature continues to rise. During this time the cardiovascular system is put under increasing stress and the flow of blood to the brain can be reduced. Unconsciousness may then result and death in the extreme case.

Questions:

1. During a warm day a walker loses 1.5 kg of perspiration by evaporation. Given that the latent heat of vaporization is 2.25 MJ/kg, calculate how much thermal energy is required to achieve this. (Answer: $3.375 \cdot 10^6$ J)

2. A person inhales in one breath $5 \cdot 10^{-4} m^3$ of dry air at atmospheric pressure and 20^0C. The air is then warmed to the body core temperature of 37^0C in the lungs. If the person takes 12 breaths per minute, calculate the heat transferred per minute to the air from the body. Assume that there are no pressure changes in the inhaled air during respiration. (Answer: $1.2 \cdot 10^2$ J/min)

3. A women whose mass is 55 kg has a metabolic rate of 9 W/kg when she is running up a hill at an angle 50 to the horizontal with a constant speed of 6 m/s. At what rate is she gaining potential energy? At what rate is she using energy? (Answer: 290 W; 500 W)

Chapter 3

Noise pollution

Noise is not often thought of as a pollutant, but unwanted sound (noise) can seriously degrade the quantity of life. The acceptance of noise by people obviously depends on the individual, but it has been legislated on levels of sound acceptable to the community. The level of noise deemed to be acceptable is dependent upon:
The type of environment: acceptable levels of surroundings noise are affected by the type of activity. A library, for example, has different requirements to those on a factory floor.
Frequency structure: different noises contain different frequencies and some frequencies are found to be more annoying than lower frequency rumbles.
Duration: a short period of high level noise is less likely to annoy than a long period.

Different people have different hearing sensitivities, but average values can be measured and provide a map of the sound that the human ear can detect. The threshold of hearing is the weakest sound that the average human hearing can detect. The threshold varies slightly with the individual, but it is remarkably low. There is also high threshold, the threshold of pain, which is the strongest sound that the human ear can tolerate.

Absolute measurements of sound intensity can be expressed in either Wm^{-2} or in sound pressure, Pa, but such units do not correspond directly to the way in which the human ear responds to sound levels. Since the human ear has a non-linear response to the energy content of sound, a logarithmic scale is used to describe the response of the ear. It is converted to sound level measured in decibels, dB

$$\mathrm{L} = 10 \cdot \log(\frac{\mathrm{I}}{\mathrm{I_0}}) \text{ or} \tag{3.1}$$

$$L = 20 \cdot \log(\frac{p}{p_0}), \tag{3.2}$$

where $I_0 = 10^{-12}Wm^{-2}$ and $p_0 = 2{\cdot}10^{-5}$ Pa are the values for the threshold of hearing, I and p the intensity and pressure of the sound being measured. The faintest audible sound (at 1000 Hz) is rated as 0 dB. Normal speech is 50 dB, road traffic 70 dB and an aircraft engine at close range is about 120 dB.

Example: Noise pollution.

If one jet causes a sound level of 120 dB on take off, what is the sound level of three such jets taking off simultaneously?

Solution:

For one jet: L = 120 dB gives us $I = 1Wm^{-2}$.
For three jets: $I = 3$ Wm^{-2} giving $L = 10 \cdot \log(3/10^{-12}) = 124.8$ dB.

3.0.1 Domestic noise and the design of partitions

The sound level in a room is determined by a number of factors. If there is no internal source of noise, then these factors are:
Direct transmission: Sound can come from neighboring rooms or outside.
Flanking transmission: This occurs through walls, floors, or ceilings.
Contact noise: This involves sound transmission through floors or ceilings.

The degree of direct transmission from outside is determined by the insulating properties of the separating walls. The transmission loss (T_L) is defined as:

$$T_L = 10 \cdot \log(\frac{P_I}{P_T}), \tag{3.3}$$

where P_I is the power incident on the wall and P_T is the power transmitted through it. Another, more common measure is the noise insulation, R. This is defined as

$$R = 10 \cdot \log(\frac{I_I}{I_T}), \tag{3.4}$$

where I_I is the intensity of noise incident on a wall and I_T is the intensity transmitted. These two measures can be related. The power incident on a partition of area A is $P_I = I_I A$. The power transmitted into the receiving room is equal to the rate at which energy is absorbed in the receiving room (assuming that no energy is transmitted onwards) and, so, $P_T = I_T W$, where W is the sound absorption of the receiving room. Thus, putting these together,

$$T_L = R + 10 \cdot \log(\frac{A}{W}). \tag{3.5}$$

R is usually measured for a band of frequencies. To reduce noise level, R must be large. To achieve this, it is necessary to:
(i) use absorbing materials (such as foams),
(ii) have hollow (cavity) walls,
(iii) double glazed windows. It is necessary to avoid resonances (i.e. must not set up a standing wave in the cavity).

For a planar, non-porous, homogeneous, flexible wall it can be shown that

$$T_L = 20 \cdot \log(f\rho_A), \tag{3.6}$$

where f is the frequency and ρ_A is the mass per unit area. This is often called the mass law. This usually gives an overestimate of the transmission loss since it ignores the effect of the stiffness of the panel. The point is that the panel can support flexural waves that, above a critical frequency, can be excited by the sound waves. The incoming sound wave couples effectively to these flexural waves which transmit sound through the panel. A properly designed panel must ensure that these flexural waves (which are unavoidable) cannot couple to sound waves within hearing range.

Even if problems with direct partitions are solved (sometimes by using double partition walls), there remain two other basic difficulties:
reflected (flanking) noise. This can be reduced by:
-covering walls with absorbing materials (tapestries) or objects that break up the wave front (pictures, china ducks etc),
- internal cavity walls,
- special dishes in walls which use destructive interference to remove reflected waves,
- remove all direct paths (i.e. fit draught excluders).
Contact noise. Reduced by using sprung floors and placing vibrating equipment on shock absorbers like rubber mats. **Questions:**

A washing machine generating 90 dB of noise is turned on at the same time as a ghetto-blaster generating 100 dB is on in the room. What is the total noise level in the room? (Answer: 100.4 dB)

Chapter 4

Atmosphere and radiation

4.1 Structure and composition of the atmosphere

The Earth's atmosphere, held by gravity, forms a gaseous envelope around the planet. Its density diminishes rapidly with altitude, with 90% of the mass concentrated within the first 20 km and 99.9% within the first 50 km. As altitude increases, the atmosphere becomes increasingly sparse, eventually merging with interstellar space. Relative to Earth's radius, the atmosphere constitutes a thin ring, underscoring its strong anisotropic properties.

Temperature decreases with altitude at a rate of 6 K/km (up to about 15 km), while horizontal temperature gradients, associated with weather phenomena, are approximately 0.05 K/km. The atmosphere comprises distinct horizontal layers delineated by temperature, each termed a sphere, with boundaries called pauses. These layers include:

• Troposphere (0-10 km). This is the lowest and contains 80% of the mass. Almost all the weather is confined to the troposphere. In particular, it contains the clouds. The temperature falls linearly with height until at the top of the troposphere the temperature is approximately –500C.
• Stratosphere (10-50 km). Above the tropopause the temperature begins to rise again until at about 50 km the temperature is about +100C. The upper part of the stratosphere contains ozone – an essential molecule for life on Earth since it filters out (harmful) UV radiation.
• Mesosphere (50-85 km). Above the stratopause the temperature falls rapidly to

about –800C. This is the coldest region of the atmosphere.

• Ionosphere (100-200 km). This is an intensely ionised region of the atmosphere and the temperature rises rapidly. Solar UV ionises the molecules of the atmosphere

$$\mathrm{h}\nu + \mathrm{AB} \rightarrow \mathrm{AB}^{+} + \mathrm{e}^{-} \tag{4.1}$$

The ionosphere reflects radio waves and is also the region of the aurorae seen in high latitudes in the northern (aurora borealis) and southern (aurora australis) hemispheres. Intense visible and UV lines caused by electron (or proton) collisions

$$\mathrm{e}^{-}(\mathrm{energy\,E_i}) + \mathrm{AB} \rightarrow \mathrm{AB}^{*} + \mathrm{e}^{-}(\mathrm{energy\,E_f}) \tag{4.2}$$

where AB can be O_2 or N_2. The energy $\Delta \mathrm{E} = \mathrm{E_i} - \mathrm{E_f}$ given up by the electron excites the molecule to the AB$*$ state. This then decays back to the ground state, releasing a photon of frequency $\nu(h\nu = \Delta E)$.

• Thermosphere (200-500 km). Temperature rises rapidly and varies strongly with the time of day, degree of solar activity and latitude. Variation between 4000C and 2000^0C possible. Minimum temperatures are at sunrise, and maximum at about 1400^0C. However, the pressures are very low, and there is little heat transfer. The idea of a temperature is becoming increasingly meaningless (high vacuum). Better to think in terms of molecular speeds.

• Exosphere (500- about 1000 km). Atoms and molecules are sparse and can escape into space.

• Magnetosphere (above 1000 km). In this region the Earth's magnetic field interacts with the solar wind and traps charged particles (electrons and protons) in the, so called, Van Allen belts.

4.1.0.1 Residence time

The residence time (τ) denotes the average lifetime of a gas molecule in the atmosphere, which is equally significant as its concentration, particularly in the context of pollutants. It's calculated as the ratio of the total average mass of the gas in the atmosphere ($<$m$>$) to the total average influx or out-flux ($<$F$>$). The reciprocal of τ represents the turnover rate of the gas. A smaller τ implies a shorter duration for the molecule or atom in the atmosphere.

For instance, molecules with small τ values tend to be reactive and may not distribute uniformly throughout the atmosphere, leading to localized effects like acid rain. Alternatively, these compounds might participate in cycles, such as the hydrologic cycle for water, where water circulates from sea to clouds to rain and back to the sea with a residence time of approximately 10 days.

Based on residence time, atmospheric constituents can be classified into three categories:

1. Permanent: These gases have very long τ values, on the order of a few million years, examples include N_2, O_2, and rare gases like CO_2 (though there may be exceptions).
2. Semi-permanent: These gases have τ values ranging from months to years, including CH_4, N_2O, CO, and CFCs.
3. Variable: Gases with τ values ranging from days to weeks, including ozone (O_3) involved in the stratospheric cycle, water vapor (H_2O) in the tropospheric cycle, as well as pollutants like SO_2, H_2S (associated with acid rain), NO_2, NH_3 (from car exhaust), and those involved in the nitrogen cycle.

4.1.0.2 Photochemical pollution

Photochemical pollution encompasses both primary and secondary pollutants. Primary pollutants are directly emitted chemical species, while secondary pollutants form through local chemical reactions involving primary pollutants. Often, the most significant damage arises from secondary pollutants rather than primary ones. For instance, while sulfur dioxide may be emitted from a power station, its conversion into sulfuric acid is more harmful to the local environment.

Urban areas face major hazardous pollutants including carbon monoxide, nitric oxide, sulfur dioxide, ozone, particulate matter, and smog. Photochemical smog, a new form of smog, is triggered by sunlight and involves the following series of reactions:

1. $NO_2 + \gamma(\lambda < 385nm) \rightarrow NO + O$
2. $O + O_2 + \text{(any molecule)} \rightarrow O_3 + \text{(any molecule)}$
3. $O_3 + NO \rightarrow O_2 + NO_2$

The molecule in the second step can be any species, serving to absorb the reaction energy. While there's no net change in the concentration of reactants, these reactions convert radiation energy into molecular kinetic energy. However, in a steady state, intermediate species maintain finite concentrations. Ozone concentration typically ranges from 0.005 to 0.05 particles per million (ppm) near the surface, and it's poisonous itself. Moreover, ozone reacts with olefins present in car exhaust, such as ethylene and propylene, forming irritating compounds like formaldehyde, acrolein, and peroxy-acetylnitrate (with NO2), causing eye and nose irritation. Ozone, produced by photocopiers as well, is recognizable by its smell. Exposure to ozone levels of 80-120 particles per billion (ppb) for a few hours can lead to respiratory problems.

4.1.0.3 Atmospheric aerosol

Atmospheric aerosols consist of suspended solid or liquid particles, such as dust particles, and originate from various sources:

- Combustion processes like forest fires or industrial activities produce soot particles.
- Gas-phase reactions contribute particles like sulfates or nitrates.
- Natural phenomena such as wind and water erosion of rocks disperse solid particles.
- Sea-spray releases salts into the atmosphere.
- Volcanic eruptions emit volcanic ash and particles.

Typical concentrations of atmospheric aerosols range from 10^3 particles per cubic centimeter (cm^{-3}) over the ocean to 10^5 particles cm^{-3} over cities. These particles vary in size, ranging from aggregates of a few hundred molecules (about 1 nanometer in diameter) to larger particles of about 10 micrometers.

The removal of aerosols from the atmosphere depends on their size:

- Aitken nuclei, with diameters less than 0.1 micrometers, are predominant in nucleating cloud droplets. They comprise about 20% of the aerosol mass and are rapidly reduced in number through collisions and coagulation.
- Large nuclei (0.1-1 micrometer in diameter) make up approximately 50% of the aerosol mass. While less numerous, they are excellent cloud droplet nuclei.
- Giant nuclei (greater than 1 micrometer in diameter) represent at least 30% of the aerosol mass. These nuclei, mainly generated by fine dust lifted by arid winds, are effective droplet nuclei and are removed from the atmosphere by falling rain. Gravitational settling becomes significant when updrafts are low.

4.2 Atmospheric pressure

In the context of climbing mountains, both pressure and temperature decrease with increasing height. The pressure in the troposphere decreases exponentially with height according to the expression:

$$p = p_0 \cdot e^{-\frac{gh}{RT}}$$

Here, p_0 represents the atmospheric pressure at the surface (at $h = 0$ m), and p represents the pressure at a height h. This exponential decrease in pressure means that 90% of the mass of the atmosphere is contained within the first 21 km, and 99.9% within the first 50 km. Consequently, pressure decreases from 105 Pa at the Earth's surface to 104 Pa at 20 km, and further to 102 Pa at 50 km.

Additionally, temperature decreases with altitude, a phenomenon referred to as the lapse rate ($\frac{dT}{dz}$). This relationship between pressure, temperature, and altitude plays a significant role in understanding atmospheric behavior and is crucial in various scientific and practical contexts.

4.3 Escape velocity

The Earth's atmosphere is held in place by gravity, and the concept of escape velocity helps us understand the minimum speed needed for an object to break free from Earth's gravitational pull. When a rocket of mass m is launched from Earth's surface with a velocity just sufficient to escape Earth's gravity, the work done is given by:

$$W = m \cdot \left(\frac{\gamma M}{R}\right)$$

Here, M and R represent the mass and radius of the Earth, and γ is the gravitational constant. To escape, the kinetic energy of the rocket must balance the work done in overcoming Earth's gravitational potential, expressed as:

$$\frac{1}{2}mv^2 = \frac{\gamma M}{2R}$$

From this equation, we find the escape velocity v to be:

$$v = \sqrt{\frac{2\gamma M}{R}}$$

Since $g = \frac{\gamma M}{R^2}$, the escape velocity is represented as $v = \sqrt{2gR}$. Substituting Earth's values ($g = 9.81\,\mathrm{m/s}^2$ and $R = 6400\,\mathrm{km}$), we find $v = 11.2\,\mathrm{km/s}$.

This analysis applies equally to molecules in the atmosphere. The distribution of speeds of molecules follows the Maxwell speed distribution. At a temperature of 288 K, the most probable speeds for O_2 and N_2 molecules are 387 m/s and 414 m/s respectively, which are well below the escape velocity. However, for lighter gases like He and H_2, with most probable speeds of 1094 m/s and 1550 m/s respectively, there is a finite probability that they may exceed the escape velocity.

Over billions of years, lighter gases like He and H_2 have been lost to space due to their ability to achieve escape velocity and the higher temperatures at the top of the atmosphere, facilitating their escape.

4.4 Ozone

Ozone, though a minor constituent of the Earth's atmosphere, comprising only 0.2% of its mass, plays a pivotal role in supporting both plant and mammalian life. It acts as a protective shield against harmful solar ultraviolet (UV) radiation, absorbing wavelengths shorter than 293 nm and preventing them from reaching the Earth's surface. This shielding effect is crucial for maintaining the health of ecosystems and protecting living organisms from the detrimental effects of UV radiation.

The presence of ozone in the atmosphere is unique among atmospheric molecules, as it possesses a strong absorption band between 210 and 300 nm. Consequently, ozone efficiently filters out the Sun's ultraviolet radiation below 300 nm, safeguarding life on Earth.

Ozone is primarily found in a thin layer known as the ozonosphere, with maximum concentrations observed between 20 and 26 km above the Earth's surface. Its formation occurs through a series of photochemical reactions initiated by the photodissociation of molecular oxygen (O_2) by solar radiation with wavelengths below 175 nm:

$$h\nu(\lambda < 175 \text{ nm}) + O_2 \rightarrow O + O$$

The resulting free oxygen atoms can then combine with oxygen molecules in a three-body collision process, involving an additional atom or molecule capable of absorbing excess energy, to form ozone:

$$O + O_2 + M \rightarrow O_3 + M$$

Most ozone is generated in equatorial regions where solar UV light is abundant, and it is subsequently transported towards the poles, leading to significant seasonal fluctuations in ozone concentrations. Natural processes, such as photo-dissociation and collisional dissociation, contribute to the maintenance of ozone equilibrium in the atmosphere:

$$h\nu + O_3 \rightarrow O_2 + O$$

$$O + O_3 \rightarrow 2O_2$$

However, ozone equilibrium is delicate and susceptible to disruption by external factors. The presence of man-made pollutants, transported from the troposphere, can disturb this equilibrium, leading to a rapid decrease in ozone concentrations.

Furthermore, natural destruction mechanisms, such as the presence of the hydroxyl radical (OH) formed by the photo-dissociation of water vapor, contribute to ozone depletion:

$$\mathrm{OH} + O_3 \rightarrow \mathrm{HO}_2 + O_2$$

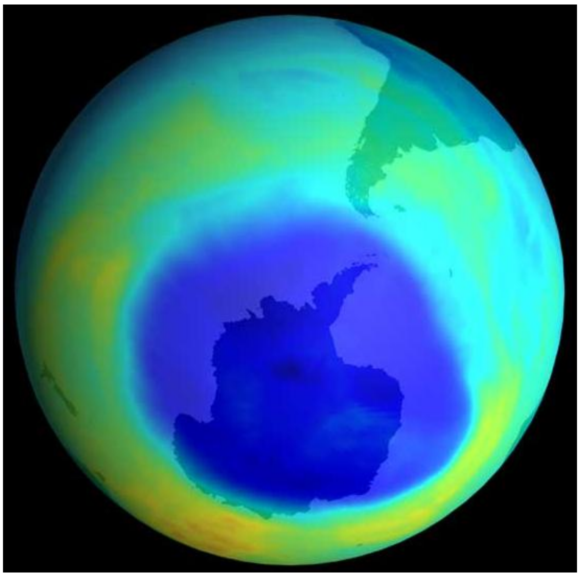
$$\mathrm{HO}_2 + O \rightarrow \mathrm{OH} + O_2$$

$$\text{Net: } O + O_3 \rightarrow 2O_2$$

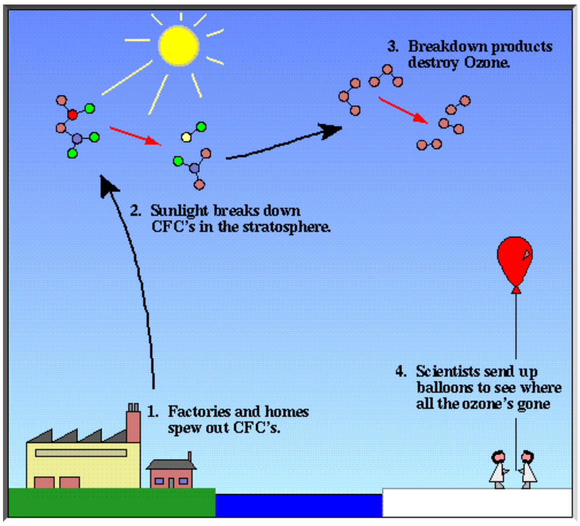
These processes explain why ozone is primarily found at altitudes between 10 and 50 km, where conditions favor its stable equilibrium. However, the introduction of new chemicals into the stratosphere can disrupt this balance, highlighting the fragility of the ozone layer and the importance of conservation efforts in preserving environmental and human health.

4.4.0.1 Ozone hole

The ozone hole, discovered in 1985 above Antarctica, represents a significant reduction in ozone concentrations, which has been observed to expand and deepen over time. The explanation for this alarming ozone depletion was initially proposed by



Molina and Rowland in 1994. They suggested that the widespread use of chlorofluorocarbons (CFCs), such as CF_nCl_{4-n}(typically with $n = 2$), in refrigerators and aerosol cans contributed to the catalytic destruction of ozone. While chemically stable in



the troposphere, CFCs can undergo breakdown in the stratosphere when exposed to solar radiation, releasing chlorine atoms. These chlorine atoms then participate in catalytic cycles, forming ClO species:

1. $ClO + O_3 \rightarrow ClO_2 + O_2$

2. $ClO_2 + O \rightarrow ClO + O_2$

The net effect of these reactions results in the destruction of ozone molecules:

$$O + O_3 \rightarrow 2O_2$$

Remarkably, ClO molecules are conserved in this process, making the mechanism far more destructive than naturally occurring mechanisms involving the hydroxyl radical (OH). A single ClO radical can destroy several hundred ozone molecules before it is eventually removed from the ozone chemical cycle.

This explanation sheds light on the concerning phenomenon of ozone depletion, highlighting the detrimental impact of human activities, particularly the release of CFCs, on the delicate balance of ozone in the Earth's atmosphere.

4.4.0.2 Ozone in polar region

The appearance of ozone holes in the Polar Regions, particularly over Antarctica, can be attributed to the unique wind systems in these areas. The polar winter vortex, characterized by its stability, is especially prominent due to the expansive landmass of Antarctica. The extremely cold temperatures, reaching as low as 183 K at heights of 15 km, create conditions conducive to the formation of ice crystal clouds, which serve as catalysts for ozone-depleting reactions. These heterogeneous reactions are highly efficient, facilitated by the presence of ice crystals, which eliminate the need for complex three-body collisions required in gas-phase reactions.

Chlorofluorocarbons (CFCs), released into the atmosphere, become trapped within the polar vortex and undergo concentration. As the polar spring arrives and solar radiation returns to illuminate these latitudes, the vortex begins to dissipate, releasing the concentrated CFCs. This release triggers ozone depletion as the air warms and the vortex breaks down. Ozone loss has also been observed at high northern latitudes, although the mechanisms may differ, with the Arctic vortex being smaller and potentially involving different ozone-depleting processes, such as the role of stratospheric aerosols and bromine compounds.

An unexpected consequence of ozone depletion is the cooling of the stratosphere. Ozone, by absorbing ultraviolet (UV) radiation, plays a crucial role in heating the stratosphere. Consequently, the loss of ozone results in a cooling effect on the stratosphere. This phenomenon partially offsets the effects of increased greenhouse gases, contributing to the complex interplay of atmospheric dynamics and climate change.

4.5 Terrestrial radiation

Not all solar radiation incident on Earth is absorbed by the atmosphere or the surface. A significant portion is reflected back into space from various surfaces, contributing to what is known as the planetary albedo, typically denoted by the symbol a. Here's a rough breakdown of Earth's energy budget:

- 33% of the solar flux is reflected back into space, primarily from cloud tops (26%), but also from the ground (2.5%) and from dust and aerosols in the atmosphere (4.5%).
- 22% of solar radiation is absorbed by the atmosphere, including 3% absorbed by clouds.
- 32.5% of solar radiation is scattered by the atmosphere. Of these, 28% subsequently reach the ground, while 4.5% are scattered back into space, contributing to the planetary albedo.
- 17% of solar radiation reaches the ground directly. Of these, 14.5% are absorbed, and 2.5% are reflected back into space.

It's important to note that 45% of solar radiation eventually reaches the Earth's surface in one way or another. The annual average rate of solar energy input per unit horizontal area at the top of the atmosphere is approximately 336 W/m^2. Therefore, the solar input to the Earth's surface is around 151 W/m^2. If all of this energy were fully absorbed, it would rapidly heat the Earth's surface to an intolerable temperature.

Thus, the rate of absorption of solar energy must be balanced by a re-emission of this energy until a steady state is achieved. This principle applies equally to the 19% absorbed by the atmosphere. The Earth effectively converts solar visible and UV radiation into terrestrial infrared radiation, which is emitted back into space. This process helps maintain Earth's energy balance and temperature equilibrium.

4.6 Earth as a black body

The Earth, including its clouds, behaves as a black body with an effective temperature T_E. The Sun, emitting as though it were a black body with a surface temperature of approximately 6000 K, radiates solar energy onto the Earth. Assuming the Earth emits terrestrial radiation as a spherical black body of radius R_E and temperature T_E, the total power output of the planet is given by the Stefan-Boltzmann law as

$P = 4\pi R_E^2 \sigma T_E^4$.

The rate of absorption of radiation from the Sun is $S(1-a)\pi R_E^2$, where a is the albedo (the integral absorptivity of the Earth) and S is the solar constant. These rates of absorption and emission must balance, leading to the equation:

$$4\pi R_E^2 \sigma T_E^4 = S(1-a)\pi R_E^2$$

From this equation, we can evaluate the effective temperature of the Earth as:

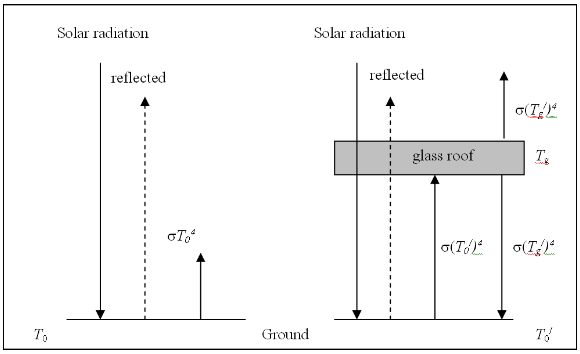
$$T_E = \left(\frac{S(1-a)}{4\sigma}\right)^{\frac{1}{4}}$$

It's noteworthy that T_E is independent of the radius of the Earth. Considering $a = 0.31$ and $S = 1353\,\mathrm{Wm^{-2}}$, we find $T_E = 255\,\mathrm{K}$ (-18°C). However, this temperature is notably colder than the mean surface temperature of the Earth, which is 288 K (15°C). This discrepancy arises because we have neglected the atmosphere entirely, including internal energy transfers between the atmosphere and the surface. In essence, we have focused solely on the interface between the atmosphere and space while disregarding the interface between the atmosphere and the ground.

4.6.1 Greenhouse effect

The elevation of the Earth's surface temperature from -22°C to +15°C is primarily attributed to the greenhouse effect, a mechanism akin to how a greenhouse operates.

The atmosphere, in this analogy, functions as a greenhouse roof. Initially, without considering the greenhouse effect, the Earth's temperature is notably lower than observed. Introducing a "glass roof" above the ground allows solar radiation to



penetrate while inhibiting the escape of infrared radiation into space. Consequently, the roof heats to a characteristic temperature T_g and radiates both downward to the ground and upward into space. As a result, the ground receives more energy than before, causing its temperature to rise until a new equilibrium is attained, where both the ground and the "roof" emit as much as they absorb.

In this equilibrium state, the upward emission from the glass roof must match the emission from the ground without the roof, implying $T_0 = T_g^/$. Considering the energy balance, the net energy received by the ground from the sun per square meter must be σT_0^4, and the energy radiated back from the glass roof must be σT_g^4. Thus, the new ground temperature is determined by $T_0^4 = T_g^4$, leading to $2\sigma T_0^4 = \sigma(T_0^/)^4$. Consequently, the new ground temperature becomes $T_0^/ = T_0^{1/4} = 298\,\mathrm{K}$ (25°C).

However, this temperature is higher than observed because the model of the atmosphere employed is overly simplistic. It assumes the glass roof only permits solar radiation to pass through while entirely blocking terrestrial radiation from the surface. Moreover, it treats the glass roof as another black body, simplifying the balance of upward and downward radiation. In reality, the balance is more complex due to the atmosphere's actual behavior.

4.6.1.1 Greenhouse gases

Greenhouse gases play a crucial role in trapping infrared radiation emitted by the Earth's surface, particularly in the wavelength range of 5-25 μm, while allowing visible light to pass through. Molecules exhibiting this behavior include water vapor (H_2O), carbon dioxide (CO_2), and ozone (O3). Water vapor absorbs radiation in bands below 4 μm, with intense absorption at 6.3 μm and a strong band beyond 9 μm. Carbon dioxide has a robust absorption band at 13-17 μm, while ozone absorbs in both the visible and infrared regions, notably with an intense narrow band at 9.7 μm, though it's primarily relevant in the stratosphere.

The crucial components of the "greenhouse roof" are carbon dioxide and water vapor, given their significant contributions to trapping infrared radiation. The effectiveness of a gas in enhancing the greenhouse effect is quantified by its global warming potential (GWP), which measures the added surface warming per unit molecule compared to CO_2. GWPs are expressed in terms of the number of CO_2 equivalent molecules.

4.6.2 Global warming

Global warming, driven largely by human industrial activities, has led to significant alterations in Earth's radiation balance by intensifying the natural greenhouse effect. The increase in atmospheric carbon dioxide (CO_2) levels, primarily due to industrial emissions, has seen a notable rise from 280 parts per million (ppm) to 350 ppm since 1770, marking a 25% increase, with the most significant uptick observed in the last 50 years.

Climate modeling efforts have focused on understanding the atmosphere's dynamics and the impact of changing parameters like CO_2 concentrations and solar flux. These models are complex, integrating factors such as atmospheric dynamics, equation of state of gases, thermodynamics, cloud effects, convection, ocean-atmosphere coupling, long-term climate systems, and terrain variations.

CO_2 isn't the sole greenhouse gas contributing to global warming. Methane (CH4) and nitrous oxide (N_2O) also play substantial roles, with methane being 7.5 times more effective at trapping heat than CO_2. Additionally, chlorofluorocarbons (CFCs) and ozone (O3) contribute to warming, albeit indirectly. The combined impact of these gases since 1800 has resulted in CO2 contributing 55%, CH4 15%, CFC-12 21%, N_2O 4%, tropospheric O_3 2%, and other gases 3% to global warming.

Climate models suggest that as CO2 concentrations rise, global average temperatures will increase by a few tenths of a degree over the next few years and by approximately 1.5°C over the following 70 years. This gradual increase, influenced by oceanic thermal inertia, will lead to significant climatic changes, including desertification in some regions and increased rainfall in others, impacting agriculture and ecosystems.

One prominent consequence of global warming is rising sea levels, driven by thermal expansion and melting ice. Thermal expansion, particularly in warmer waters, can lead to substantial sea-level rise, with predictions ranging from 20-50 cm over the next century, posing significant challenges to coastal areas worldwide.

Questions:

• If carbon monoxide (CO) is in the air at 1 ppm (particle per million) what are the number of CO in 1 m^3, and what is the mass of CO in 1 m3? (Answer: 2.7·1019 molecules per m^3; 1.25 mg/m^3)

• If the ratio of 4He to 40Ar entering the Earth's atmosphere from radioactive decay during the past 109 years is close to unity, why is the present-day 4He measured in the Earth's atmosphere much less than that of 40Ar?

• Discuss why ozone is produced mainly in the equatorial regions and hence how it is circulated around the Earth.

• Why is ozone restricted to such thin layer in the Earth's atmosphere between 20 and 50 km?

• Explain why it is imperative that there are greenhouse gases in the Earth's atmosphere.

• Calculate the density of air at the summit of Mount Everest where the pressure and temperature are $3.13 \cdot 10^4$ Pa and -38.5^0C, respectively. (Answer: 0.465 kg/m^3)

• More than 2/3 of the Earth's surface is covered by water. Solar radiation reaches the Earth as roughly parallel beams of mean intensity $1.4kW/m^2$. Of this radiation, on average 50% reaches the surface. The oceans have a mean reflectivity of 7%, while they reradiate 35% of the energy they absorb. using these data, estimate the mean annual rainfall over the whole of the Earth's surface, expressed as a depth in milimetres. (Answer: 1000 mm)

• The value of the solar constant can be found using Stefan's law. Use the data: $\sigma = 5.6710 - 8Wm^{-2}K^{-4}$, TSun = 5770 K, RSun = 6.98·108 m, REarth's orbit = 1496·108 m) to calculate its value. (Answer: 1368 W/m^2)

Chapter 5

Water

The density of water is at a maximum at $4^0 C$. Ice I (there are at least ten phases of ice) is an open structure, held together by hydrogen bonds, and is less dense than liquid water. This is useful since the oceans freeze from the top down, not the bottom up. The latent heat of fusion (i.e. the heat required to convert 1 kg of ice to water without changing the temperature) is 0.334 kJ/kg. The latent heat of vaporisation (i.e. the heat required to convert water to vapour without a change of temperature) is 2300 kJ/kg.

5.1 Hydrosphere

The total volume of water on the Earth is about 1284 M km^3; 97% of this is in the oceans. If it were spread evenly over the Earth, the planet would be covered to a depth of 2.8 km; 2.25% is locked up in the polar ice-caps and in glaciers; about 0.75% is in soil, lakes and rivers; 0.035% is in the atmosphere. To give some idea of what this amount is: if all the water vapour in the atmosphere were instantly converted to rain, the total rainfall (averaged over the Earth's surface) would be about 3 cm. Yet the annual average rainfall here is 90-100 cm. There is a hydrologic cycle.

5.2 Hydrologic cycle

Water is cycled between the oceans and the atmosphere. As always, the cycle is driven by the Sun. Most of the water vapour in the atmosphere (84%) comes from the oceans. Transpiration from plant leaves accounts for most of the rest. The Sun heats the water in the oceans (and on the land surface), giving evaporation. The warm, moist air rises, expands (under reduced pressure higher in the atmosphere) and cools. The water vapour condenses to form clouds. The winds then carry the clouds across the Earth's surface until the water is released as precipitation (rain, hail or snow) to fall on the Earth for further recycling. Most of the precipitation falls into the oceans (75% of the surface of the planet being ocean). The rate of circulation of water within the hydrologic cycle is very rapid. Since the total mass of water in the atmosphere is constant, precipitation is balanced by evaporation. Thus, comparing throughputs shows that the average residence time of water molecules in the atmosphere is about 10 days.

5.3 Water in the atmosphere

Water significantly influences atmospheric dynamics across several domains:

Thermodynamics: Water's presence drives condensation and evaporation processes.
Cloud Formation: Water vapor contributes to cloud formation, affecting planetary albedo and precipitation patterns.
Atmospheric Cleansing: Rain removes substances like hygroscopic aerosols and soluble gases, purifying the atmosphere.
Chemistry: Water serves as a solvent and participates in atmospheric reactions.
Radiation Absorption: As a major greenhouse gas, water absorbs radiation.

Water vapor content in the atmosphere is constrained by saturation vapor pressure, the partial pressure of water vapor in equilibrium with the condensed phase. The Clausius-Clapeyron equation describes this equilibrium, relating partial pressure (p) to temperature (T) and latent heat of evaporation (L):

$$\frac{\mathrm{d}\ln \mathrm{p}}{\mathrm{dT}} = \frac{\mathrm{L}}{\mathrm{RT}^2} \tag{5.1}$$

Latent heat (L) of vaporization can be expressed as $\mathrm{L} = \mathrm{L_S}/\mathrm{M_v}$, where $\mathrm{L_S}$ is the latent heat and $\mathrm{M_v}$ is the molecular weight. Water's specific gas constant $\mathrm{R_{S(H_2O)}} = \frac{1000\mathrm{R}}{\mathrm{M_v}}$.

The saturated vapor pressure (e_s) is strongly temperature-dependent, expressed by:

$$\mathrm{e_s} = \mathrm{e_s^0} \exp\left(\frac{-\mathrm{L}}{\mathrm{R}}\left[\frac{1}{\mathrm{T}} - \frac{1}{\mathrm{T_0}}\right]\right) = \mathrm{e_s^0} \exp\left(\frac{-\mathrm{L_S}}{\mathrm{R_{S(H_2O}}}\left[\frac{1}{\mathrm{T}} - \frac{1}{\mathrm{T_0}}\right]\right) \tag{5.2}$$

Here, $\mathrm{e_s^0}$ represents the saturated vapor pressure at standard temperature and pressure (S.T.P.), and T is the temperature.

Relative humidity (RH) indicates the ratio of water vapor's partial pressure to the saturated vapor pressure at a given temperature, expressed as a percentage:

$$\mathrm{RH} = 100 \cdot \frac{\mathrm{e(T)}}{\mathrm{e_{s(T)}}} \tag{5.3}$$

For instance, in tropical regions (25^0C), water vapor partial pressure reaches 32 mb, contrasting with polar vortex conditions (-20°C) where it diminishes to 1.2 mb.

5.4 Clouds

Clouds are categorized into various types based on their characteristics:

Cirrus (cirro): High-altitude clouds composed of ice crystals, often appearing as white bands or filaments.
Stratus (strato): Layer clouds, either continuous or showing structured patterns.
Alto: Middle-height clouds.
Cumulus (cumulo): Clouds with vertical circulation, resulting in a fluffy appearance distinct from stratus clouds.
Nimbus (nimbo): Clouds producing precipitation, such as rain, snow, or hail.

These classifications can combine to form different cloud types, such as cirrocumulus, cirrostratus, altocumulus, altostratus, nimbostratus, stratocumulus, stratus, cumulus, and cumulonimbus clouds.

For instance, a cumulonimbus cloud, a type of thunderstorm cloud, exhibits large vertical air movement and typically produces heavy rainfall.

Considering an average cumulus cloud, which has a roughly cylindrical shape with dimensions around 2 km in diameter and 2 km in depth, its volume is approximately $6.28 \times 10^9 \mathrm{m}^3$. A typical cumulus cloud contains between 50-500 million water droplets per cubic meter, with each droplet having a radius of about 10μm. The mass of a single droplet, assuming the lowest reasonable density, is around 4.2×10^{-12} kg. Thus, the total mass of the cloud is approximately 1.3×10^{6} kg.

However, if all the water in the cloud fell at once, the depth of water would be only 4.2×10^{-4} m, or 0.42 mm, which is relatively small. During a storm, numerous clouds must pass over an area to produce significant rainfall.

5.4.1 Physics of cloud formation

Cloud formation is initiated when hot, humid air rises and undergoes adiabatic expansion, leading to cooling. As the air ascends, it eventually reaches saturation, at which point excess water vapor begins to condense onto tiny particles known as cloud condensation nuclei (CCN). The process of condensation requires a partial vapor pressure higher than the thermodynamic vapor pressure due to the curved surface of the droplets. However, the presence of dissolving materials in the droplets reduces the vapor pressure. Both of these effects are dependent on the radius (r) of the droplet.

Cloud condensation nuclei originate from various sources, including natural ones such as dust from wind erosion and sea salt, as well as industrial emissions like sulfate particles and coal dust. It's estimated that around 100 such particles per cubic centimeter are necessary to initiate cloud formation.

The concentration of cloud condensation nuclei varies depending on location, with approximately 10^3 particles/cm^3 over the sea, 10^4 particles/cm^3 over land, and 10^5 particles/cm^3 over cities. Despite seemingly adequate concentrations of particles for cloud formation, the process is more complex than mere particle abundance suggests.

5.4.1.1 Growing droplets in cloud

Cloud droplets grow through condensation and coalescence mechanisms within the cloud environment. Here's a detailed explanation of the processes involved:

1. Condensation Growth:
- As water vapor condenses onto the surface of a droplet, latent heat is released, causing the droplet's temperature to rise.
- The increased temperature alters the vapor pressure around the droplet.
- A gradient of partial pressure drives a flux of water vapor towards the droplet.

- The rate of heat loss equals the heat gained by the droplet from the latent heat of water.
- This leads to a growth rate equation: $\frac{dr}{dt} = C/r$, where C is a constant.
- Integrating this equation yields $r^2 = r_0^2 + 2Ct$, where r_0 is the initial droplet size.
- However, detailed calculations suggest that it would take 1-4 hours for a droplet to grow from 2 to 30μm, whereas clouds contain many droplets larger than 10μm even though their lifetimes can be as short as ten minutes.
- Additionally, as the cloud rises, the temperature drops, causing the saturated vapor pressure to fall and super-saturation in the cloud to rise. This allows more nuclei to act as condensation sites.
- However, condensation alone cannot fully explain cloud growth.

2. Coalescence Mechanism:
- Coalescence involves the collision of two droplets to form a larger droplet.
- The probability of coalescence depends on the size of the droplets and their relative velocity.
- Droplet velocity is influenced by a balance between gravity and frictional forces.
- Terminal velocity, determined by Stokes' Law, depends on droplet size and medium viscosity.
- The rate of growth is controlled by the equation $\frac{dr}{dt} = \frac{Ewv}{4\rho}$, where E is the collection efficiency, w is the volume swept out by the falling droplet, v is the droplet velocity, and ρ is the density of water.
- There's a growth barrier at around 20μm where neither condensation nor coalescence is efficient.
- The mechanism overcoming this barrier is not fully understood and might involve increasing collision efficiency through turbulence effects or droplet-droplet interactions by electrical forces.
- Droplets with diameters of 2-3 mm are often broken apart by collisions rather than built up, and droplets larger than 6 mm become unstable due to surface tension.
- Once droplets grow beyond the cloud's ability to maintain them, they begin to fall. Only larger drops, formed by coalescence, can reach the ground as rain.
- In colder clouds, precipitation may start as ice or hail, which may or may not melt before reaching the ground.

5.4.2 Thunderstorms

Thunderstorms are atmospheric phenomena characterized by the following processes:

1. Formation:
- Thunderstorms develop when warm, moist air near the ground becomes buoyant and rises, forming small cumulus clouds.
- These clouds grow and ascend, combining to create cumulonimbus clouds with an anvil-shaped top that extends up to 1 km into the stratosphere.
- Within cumulonimbus clouds, vigorous movements such as updrafts and downdrafts occur at speeds of tens of meters per second.

2. Charge Separation:
- Movement within the clouds causes separation of electrical charges, with positive charge accumulating at the top and negative charge at the bottom.
- Charge carriers include electrons, molecules, aerosol dust, hailstones, and snowflakes.
- The process of charging involves frictional contact, freezing, melting, and break-up of water droplets.

3. Lightning:
- Once charge separation occurs, an electric field is established, and the air may become ionized, transforming a region of the atmosphere from a good insulator to a highly conducting path.
- Lightning is a high-voltage spark resulting from electrical discharge, involving the flow of ten coulombs of charge across a potential difference of about 100 million volts, with an energy of approximately 1 billion joules (about 280 kWh).
- Types of lightning include ground discharges (thunderbolts), cloud lightning (sheet lightning), air discharges (streak lightning), and ball lightning (unusual and still not fully explained).

4. Thunder:
- Thunder is the sound wave produced by a lightning stroke.
- The sudden rise in pressure within the lightning channel generates an intense sound wave similar to an explosion.
- Sound travels at approximately 330 m/s, so the sound of thunder follows behind the lightning flash.
- By counting the interval between the flash and the thunder, one can estimate the

distance of the storm (roughly 1 km for every three seconds, but echoes can complicate measurements).

Questions:

1. How does the terminal velocity of a drop depend upon its radius in the range $10^{-30}\mu$m?
2. The horizontal acceleration on an air parcel of mass 1 tonne is 10^{-4}ms^{-2}. What is the net force on the air parcel? Estimate the volume of the air parcel at sea level. (Answer: 0.1 N; 1000 m^3)
3. Given that a cumulus cloud is typically 2 km deep with a similar diameter and contains 5·107 water droplets per m3 each of 10 μm radius, calculate the depth of rainfall shoul the cloud release all its water in one instant. (Answer: 0.42 mm)
4. Calculate the space charge, assuming that there are 1000 positive ions and 900 negative ions per cm^3 in the air. (Answer: 16 pCm^{-3})

Chapter 6

Wind

6.1 Measuring the wind

The Beaufort scale, devised by Francis Beaufort around 1800, provides a qualitative description of wind speed based on its effects. This scale has been refined over time and is still used today for maritime and land-based forecasts. Table ?? is a simplified modern description of the Beaufort scale for land use, along with corresponding wind speeds in meters per second (m/s):

6.2 Principal of wind formation

The atmosphere, fueled by the Sun's radiation, acts as a colossal heat engine, driving convection on both local and global scales. The unequal distribution of solar energy between the equator and the poles generates pressure variations that propel the major wind systems within Earth's atmosphere. Winds, characterized as moving masses of air, play a pivotal role in shaping weather patterns.

In meteorology, an air mass denotes a vast expanse of air spanning millions of square kilometers, possessing relatively uniform pressure and humidity levels. While an air mass dictates the general weather conditions of a region, it may not account for localized microclimates. These air masses originate from expansive high-pressure zones prevalent across specific regions of the Earth, such as the subtropical oceans year-round and the mid to high latitude continents primarily during winter. Air

Beaufort	Description	Wind Speed (m/s)
0	Calm: smoke rises vertically.	0-0.5
1	Light air: wind direction shown by smoke drift, not by vanes).	0.5-1.5
2	Light breeze: wind felt on face, leaves rustle, vanes move.	1.5-3.5
3	Gentle breeze: leaves and small twigs move, light flags lift, large wavelets at sea.	3.5-5.5
4	Moderate breeze: dust and loose paper lift, small branches move.	5.5-8.0
5	Fresh breeze: small leafy trees sway, moderate waves.	8.0-10.5
6	Strong breeze: large branches sway, telegraph wires whistle, umbrellas difficult to use.	10.5-13.5
7	Near gale: whole trees move, inconvenient to walk against.	13.5-17
8	Gale: small twigs break off, walking impeded, high waves and foam.	17-20.5
9	Strong gale: slight structural damage.	20.5-24.5
10	Storm: considerable structural damage, trees uprooted.	24.5-28.5
11	Violent storm: widespread damage.	28.5-32.5
12	Hurricane: at sea, visibility badly affected by driving foam and spray. Sea surface completely white.	>32.5

Table 6.1: Beaufort scale description with corresponding wind speeds.

spirals outward from these high-pressure centers, known as anticyclones, shaping the global wind systems. Two notable examples pertinent to Europe include:

- The Azores anticyclone: Southwesterly winds flow towards the North Pole, while the Northeast Trades head towards the equator. This warm and humid air, classified as 'tropical maritime' in mid-latitudes, influences European weather.
- The polar continental: During winter, cold and dry air from Eurasia dominates. This air mass, with its distinct characteristics, impacts weather patterns in Europe.

The interfaces separating different air masses are termed fronts. In this context, a cold front delineates the leading edge of a cold air mass, often associated with rain. As the cold front advances, it displaces the warm, moisture-laden air of the tropical maritime mass, forcing it upward. Upon ascent, the air cools, leading to precipitation.

6.2.1 Principal forces acting on air masses

If we want to understand why the winds occur, it is necessary to consider the forces that act on the air masses in the atmosphere. To any observer stationary with respect to the surface of the Earth, there are four forces acting on a parcel of air in the atmosphere:
- gravitational,
- pressure gradient,
- Coriolis fictional force, and
- frictional force.

6.2.1.1 Gravitational force

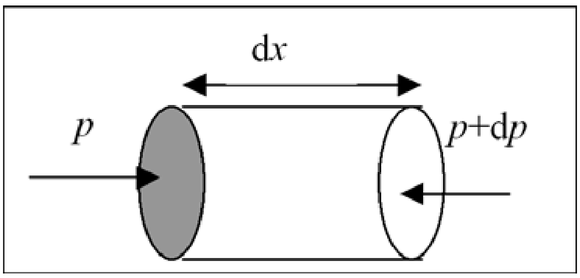
Due to the large mass of the Earth, the gravitational force is one of the strongest forces acting on the air parcel and is directed towards the centre of the Earth

$$\mathrm{F_g} = \mathrm{g}\rho\Delta\mathrm{V} \tag{6.1}$$

where g is the gravitational acceleration constant (more or less constant through the troposphere), ρ is the density of air, and ΔV is the volume of the parcel.

6.2.1.2 Pressure gradient

The pressure gradient force arises from the difference in pressure between neighboring air parcels. It represents the normal component of the force exerted by the surrounding air on a unit area of the parcel's surface. This force is inherently directed towards the parcel. In mathematical terms, let's consider a shaded cross-section with an area



dA. The net force $\mathrm{F_P}$ on the parcel due to the pressure difference can be expressed as:

$$\mathrm{F_P} = \mathrm{pdA} - (\mathrm{p} + \mathrm{dp})\mathrm{dA} = -\mathrm{dpdA} \tag{6.2}$$

Here, p represents the pressure, and dp is the change in pressure. If ρ denotes the density of air, the force per unit mass, which represents the acceleration, is given by:

$$\frac{\mathrm{F_P}}{\rho \mathrm{dAdx}} = -\frac{1}{\rho}\frac{\mathrm{dp}}{\mathrm{dA}}\frac{\mathrm{dA}}{\mathrm{dx}} = \frac{1}{\rho}\frac{\mathrm{dp}}{\mathrm{dx}} \tag{6.3}$$

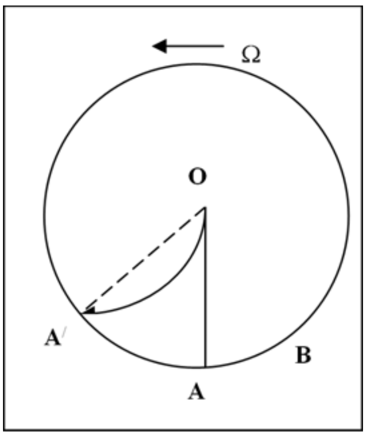
This expression shows the change in pressure with respect to distance x. By generalizing this to three dimensions, we get:

$$\tilde{\mathrm{F}}_{\mathrm{P}} = -\frac{1}{\rho}\tilde{\nabla}\mathrm{p} \tag{6.4}$$

Where $\tilde{\nabla}$ is the gradient operator.

6.2.1.3 Coriolis inertial force

The motion of air masses in the atmosphere is influenced by several forces, including the gravitational force, pressure gradient force, and the Coriolis force. The Coriolis



force is a fictitious force due to the rotation of the Earth. It appears when considering the motion of objects in a rotating reference frame, such as the Earth. To understand its effect, imagine a region around the North Pole represented as a rotating disc. As an air parcel moves horizontally away from the pole towards a point A, the rotation of the Earth causes it to follow a curved path, instead of a straight line, as viewed from the rotating frame of reference. This apparent deflection of the parcel from its intended path is the Coriolis force.

Mathematically, the Coriolis force per unit mass is given by

$$\tilde{\mathrm{F}} = -2\tilde{\Omega} \times \tilde{\mathrm{v}}_{\mathrm{g}} \tag{6.5}$$

where $\tilde{\Omega}$ is the angular velocity vector of the Earth's rotation, and $\mathbf{v}_{\mathrm{g}}$ is the velocity vector of the air parcel.

Since the atmosphere is thin compared to the Earth's radius, the wind generally blows along the local horizontal. Therefore, it's useful to split the Coriolis force into two components with respect to the local vertical. These components are given by $(\Omega \sin \phi)\,\tilde{\mathrm{z}}$ and $(\Omega \cos \phi)\,\tilde{\mathrm{y}}$, where ϕ is the latitude angle.

If the wind is moving in the y direction, then the Coriolis force is given by

$$\tilde{\mathrm{F}} = -(\Omega \sin \phi)\tilde{\mathrm{z}} \times \tilde{\mathrm{v}}_{\mathrm{g}}\tilde{\mathrm{y}} = (\tilde{\mathrm{v}}_{\mathrm{g}} \sin \phi)\,\tilde{\mathrm{x}} \tag{6.6}$$

where $\tilde{\mathrm{x}}$ and $\tilde{\mathrm{y}}$ are unit vectors in the local horizontal plane.

The Coriolis force deflects the initial movement of the air parcel until it balances the pressure gradient force, leading to a wind blowing perpendicular to the pressure gradient producing it. This balance is known as the geostrophic balance. Formally, the balance between the Coriolis force and the pressure gradient force, known as the geostrophic balance, can be expressed as:

$$\left(1 - \frac{1}{\rho}\nabla \mathrm{p} + \mathrm{f}_{\mathrm{C}}\mathrm{v}_{\mathrm{g}}\right)\tilde{\mathrm{x}} = 0 \tag{6.7}$$

where $(\mathrm{f}_{\mathrm{C}} = 2\Omega \sin \phi)$ is the Coriolis constant (approximately $10^{-4}\ \mathrm{s}^{-1}$), v_g is the air velocity at balance, and $\mathbf{x}$ represents the direction along which the pressure gradient increases (assumed to be along the x-axis). This gives

$$\mathrm{v}_{\mathrm{g}} = \frac{1}{\mathrm{f}_{\mathrm{C}}\rho}\frac{\mathrm{dp}}{\mathrm{dx}} \tag{6.8}$$

In this balance, the geostrophic wind results, and its direction is parallel to the isobars. In the northern hemisphere, the wind direction is such that lower pressure is on the left-hand side as you face downwind. Consequently, low-pressure areas in the northern hemisphere have winds rotating around them in a counterclockwise direction, which is termed cyclonic motion. Conversely, over a high-pressure area in the northern hemisphere, the geostrophic wind circulates in a clockwise direction, known as anticyclonic motion, and high-pressure weather systems are referred to as anticyclones.

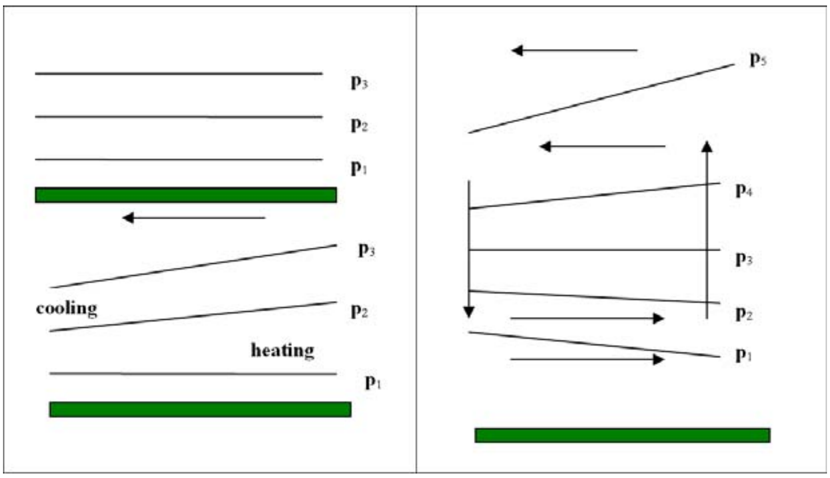
The Coriolis force is greatest at the poles and decreases as one approaches the equator, where it becomes zero. Additionally, there will be a vertical component proportional to $\cos\phi$ arising from the y vector, contributing slightly to the effective gravitational force.

In summary, the Coriolis force plays a crucial role in determining the direction of wind motion, particularly in relation to the Earth's rotation and the pressure gradient force.

6.2.1.4 Frictional force

Frictional force between the atmosphere and the Earth's surface is significant, influenced by factors like terrain features such as mountains and hills. Planting lines of tall trees serves as windbreaks to shield crops from excessive winds. The mechanics of frictional forces involve viscosity at lower altitudes and small-scale eddy mixing at higher altitudes. This layer where frictional force is crucial is termed the planetary boundary layer, with variable thickness from a few hundred meters in still air at night to 4-5 km over a hot surface with strong convection.

Understanding how wind speed varies with height is closely linked to the variability of atmospheric pressure with height, as explored in atmospheric structure discussions. Pressure differences arise from uneven heating or cooling, exemplified by an evenly temperature-distributed region where one end is heated while the other is cooled. Hot air expands while cold air contracts, creating horizontal pressure gradients. This phenomenon leads to a thermal wind, where air moves down the pressure gradient. In the warm area, the rate of pressure decrease at a fixed height corresponds to the



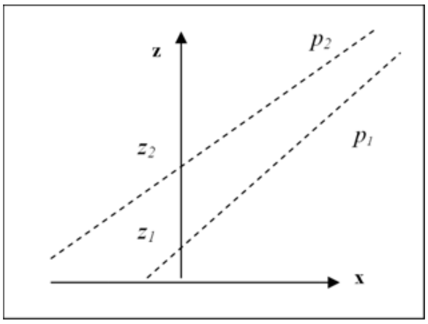
mass of air flowing out above that height, with a faster decrease in pressure at lower levels than at higher ones. Eventually, a steady state is achieved, with lower pressure in the upper part of the air column in warm areas and higher pressure in cold areas. Conversely, in the lower part of the air column, pressure is higher in cold areas and lower in warm areas. This dynamic interplay forms the basis of convection in the atmosphere.

6.2.1.5 Example: A day at the seaside.

During the day, the Sun heats the land, causing its temperature to rise above that of the sea. Since the specific heat of the land is lower than that of the sea, the air over the land becomes warmer compared to the air over the sea. Consequently, a low-level wind blows from the sea towards the land, known as a sea breeze.

At night, the land cools down below the temperature of the sea, leading to a reversal of the situation. Now, the low-level wind blows from the land towards the sea, forming a land breeze. This straightforward example illustrates one of the fundamental mechanisms driving the global circulation of the atmosphere.

We can formalize the variation of the geostrophic wind with height by considering the difference in the geostrophic wind velocity between two isobaric surfaces at different heights, z_1 and z_2. Let's denote this difference as Δv_g: Then the geostrophic wind is



along the (+y) axis and is given by:

$$v_g = \frac{1}{f_C \rho} \frac{dp}{dx} \tag{6.9}$$

We now consider the situation in the above diagram. We take two isobars, pressure p_1 and p_2 at heights z_1 and z_2, and calculate the difference in the geostrophic wind velocity between them in the upwards direction. We compute the difference Δv_g as follows:

$$\begin{aligned}
\Delta v_g &= \frac{1}{f_C}\left(\frac{1}{\rho}\frac{dp}{dx}\bigg|_{z_2} - \frac{1}{\rho}\frac{dp}{dx}\bigg|_{z_1}\right) \\
&= \frac{1}{f_C}\left(\frac{1}{\rho}\frac{dp}{dx}\bigg|_{z_2} - \frac{1}{\rho}\frac{dp}{dx}\bigg|_{z_1}\right) \\
&= \frac{1}{f_C}\left(\frac{1}{\rho}\frac{dp}{dz}\frac{dz}{dx}\bigg|_{z_2} - \frac{1}{\rho}\frac{dp}{dz}\frac{dz}{dx}\bigg|_{z_1}\right) \\
&= \frac{1}{f_C}\left(-g\frac{dz}{dx}\bigg|_{z_2} + g\frac{dz}{dx}\bigg|_{z_1}\right) \\
&= \frac{g}{f_C}\left(\frac{dz}{dx}\bigg|_{z_1} - \frac{dz}{dx}\bigg|_{z_2}\right)
\end{aligned}$$

This provides the expression for Δv_g in terms of the vertical gradients of pressure with respect to the horizontal coordinates x at the heights z_1 and z_2.

But we know that the hydrostatic equation states that

$$dp = -\rho g\, dz$$

In the diagram above, the isobars are sloped, so we can measure dp either along the x r axis or along the z axis. In the diagram above, the values of z are linked to the lines of constant pressure that intersect the x axis, so we can write

$$\frac{dp}{dx} = -\rho g \frac{dz(p)}{dx},$$

where $z = z(p)$ to remind us of the connection with the heights and the isobars. So we can therefore write as:

$$\Delta v_g = \frac{g}{f_C}\left[\frac{d\Delta z}{dx}\right].$$

We can now use the buoyancy equation, $dp = -\rho g dz$, again to determine Δz. Using the ideal gas law, $p = R_S \rho T$, and rearranging, we have

$$dp = \frac{pg}{R_S T} dz.$$

Integrating between the isobars p_1 and p_2 (and therefore between the heights z_1 and z_2) we have

$$-\int_{p_1}^{p_2} \frac{R_S T}{p} dp = \int_{z_1}^{z_2} g dz,$$

where R_S is the specific gas constant. Integrating this and assuming that g is constant, we obtain

$$-\frac{R_S T}{g} \ln \frac{p_2}{p_1} = z_2 - z_1.$$

If we replace the height-dependent temperature $T(z)$ by an average value, $\bar{T}$, then we obtain

$$\Delta z = z_2 - z_1 = \frac{R_S \bar{T}}{g} \ln \frac{p_1}{p_2}.$$

We can now differentiate this, remembering that we chose p_1 and p_2 to be constant:

$$\frac{d(\Delta z)}{dx} = \frac{R_S \bar{T}}{g} \frac{d \ln(p_1/p_2)}{dx}.$$

For the difference in the geostrophic wind velocity, we get

$$\Delta v_g = \frac{g}{f_C} \frac{d(\Delta z)}{dx} = \frac{R_S \bar{T}}{f_C} \frac{d \ln(p_1/p_2)}{dx}.$$

Finally, we can eliminate the pressure term using the hydrostatic equation to give the difference form of the thermal wind equation:

$$\frac{\Delta v_g}{\Delta z} = \frac{g}{f_C \bar{T}} \frac{d\bar{T}}{dx}.$$

As Δv_g and Δz tend to zero, the left-hand side tends to the derivative and the average temperature on the right-hand side tends to that of the isobaric surface in the middle of the layer. This gives an important relationship between the vertical variation of the geostrophic wind and the horizontal temperature gradient. The isobaric slope increases with height, so the horizontal pressure gradient also increases with height. Hence the geostrophic wind should also increase with height. The increase in velocity is in a direction that is perpendicular to the temperature gradient, with the cold region to the left and the warm region to the right of the wind vector increment (in the northern hemisphere). Thus, the geostrophic wind rotates counter-clockwise with altitude when the wind blows from a cold region to a warm one, and clockwise when the wind blows from warm to cold.

6.3 Cyclones and anticyclones

Cyclonic systems are prevalent in both low and middle latitudes, with middle latitude cyclones often associated with unfavorable weather conditions. These systems, known as depressions due to their low pressure nature, typically form as a result of the collision between cold polar air masses from the North American continent and warm tropical air masses from the Western Atlantic, influenced by the Gulf Stream. Similar interactions occur in the North Pacific and Northern Mediterranean regions.

Depressions develop when a wave forms at the boundary between these contrasting air masses. As air flows across the isobars, a low-pressure area with cyclonic motion emerges. The cyclonic motion of depressions is characterized by the movement of warm air rising above cold air, often resulting in the cold front catching up with the warm front.

In contrast, anticyclones originate over Siberia, Canada, and North Russia, dominating the winter climate of Asia and North America. These high-pressure systems generally bring dry weather but can trap low-lying clouds due to light winds. While most anticyclones last for a few days, some persist longer and are termed blocking highs, impeding the movement of cyclones and creating stationary fronts for extended periods.

Tropical cyclones, commonly known as hurricanes or typhoons, are intense low-pressure systems characterized by strong winds and heavy rainfall. They typically form over warm ocean waters, where rising humid air creates a belt of clouds around the Earth. The inter-tropical convergence zone (ITCZ), where warm and cold air masses converge, serves as a breeding ground for tropical disturbances, some of which develop into major storms.

The development of hurricanes involves a rapid fall in pressure at the disturbance's center, accompanied by rising winds forming a tight band known as the eye. As the storm matures, it moves westward within the trade winds before migrating to higher latitudes. The storm's growth and trajectory are influenced by local surface conditions, such as sea surface temperature and salinity. Hurricanes eventually decay, especially when passing over colder water or land, and are propelled by mid-latitude westerlies in higher latitudes.

These storms, occurring most frequently in late summer, contribute to the transfer of energy from the equator to the poles and are vital components of Earth's atmospheric circulation patterns.

6.4 Global convection

George Hadley's model proposed in 1735 was the first attempt to describe large-scale global convection patterns. He observed that air in lower latitudes near the equator is warmer than air in higher (polar) latitudes due to the higher solar flux received at the equator. Hadley suggested that warm tropical air rises vertically and moves poleward, while cold polar air moves equatorward. As tropical air moves northward, it loses energy through radiation and descends back to the surface, replacing the southward-moving colder air. Conversely, cold air gains heat from the warm ground and rises in equatorial regions. This circulation system transfers thermal energy from the equator to the poles, known as the Hadley cell.

However, there are notable differences between Hadley's model and real atmospheric circulation patterns. While there is indeed a low-pressure belt over the equator and a high-pressure region over the poles as predicted by Hadley, there are additional circulating cells of air. One such cell, the Ferrel cell, operates between 30°N and 60°N (and similarly in the southern hemisphere), where air rises in colder regions (at 60°N) and descends in warmer regions (opposite to the Hadley mechanism).

Additionally, there is a third cell between 60°N (and S) and the pole, known as the Polar cell, which circulates in the same direction as the Hadley cell but is much weaker.

Together, these three cells—Hadley, Ferrel, and Polar—comprise the global circulation system, transporting heat from the equator towards the poles and influencing weather patterns worldwide.

6.5 Global wind patterns

The global wind patterns are influenced by seasonal variations in atmospheric pressure across the planet. These patterns change with the shifting seasons:

1. Northern winter / Southern summer:
- Weak high pressure exists across the Arctic Ocean, with low-pressure centers southeast of Greenland and in the North Pacific.
- Between 45°N and 15°N, there's a broad region of high pressure extending from the subtropical Eastern Pacific eastwards towards India, with exceptions such as a low-pressure area in the Mediterranean.
- Weak low-pressure centers are found over the southern continents, while subtropical anticyclones dominate the main southern oceans. A belt of rapidly changing pressure gradient and strong winds exists around 40°S, followed by a high-pressure region over Antarctica.

2. Northern summer / Southern winter:
- Mid-latitude oceanic low centers in the North Atlantic and North Pacific weaken, and subtropical anticyclones shift towards the Pole.
- Extensive low-pressure systems are observed over central and southern Asia, associated with the monsoon. Similar low-pressure regions are found over subtropical areas of Africa and North America. The reverse occurs in the southern hemisphere.
- A deep low-pressure belt surrounds Antarctica.

The strongest wind patterns generated by these seasonal pressure variations are themselves seasonal:

- **Northern winter / Southern summer:**
- Mid-latitude westerlies blow from the west, primarily confined by the ocean basins.
- North-easterlies extend from the Asian anticyclone over the Arabian and South China seas.
- Trade winds blow northeast in the northern hemisphere and southeast in the southern hemisphere, converging at the Inter-tropical Convergence Zone (ITCZ).

- **Northern summer / Southern winter:**
- Mid-latitude westerlies are weaker in the northern latitudes, approximately 65% of their winter velocity.
- The ITCZ over the oceans moves north, causing a reversal of wind direction over the Arabian and South China seas, part of the monsoon.

These wind patterns are not fully explained by Hadley's model but result from the balance of forces induced by thermal convection, frictional forces of air moving across the Earth's surface, and the Coriolis force.

At higher altitudes, wind patterns are simpler due to reduced friction from continental masses. The jet streams, occurring between 30°N and 50°N (and similarly S), are significant seasonal winds. In the Southern hemisphere, the polar vortex, characterized by westerly winds high over the coasts of Antarctica, circulates in a circular pattern, preventing heat transport into the air above Antarctica. This cold air plays a role in depleting the ozone layer, explaining the observation of the ozone hole over Antarctica.

Questions:

1. Consider a low-presure system in the Southern-Hemisphere. In which direction does air circulate around the low-presure centre?
2. Using the definition for the geostrophic wind, or otherwise, explain why hurricanes do not cross the Equator. Storm force northerly winds are blowing at 25 m/s over sea area at 550N. Calculate the horizontal pressure gradient (in Pa per 10 km) associated with this speed, assuming geostrophic balance. (Answer: 35.6 P per 10 km)
3. Calculate the geostrophic wind speed for a pressure gradient of 0.03 mb/km, assuming that the Coriolis parameter f = 10-4 s-1. (Answer: 30 m/s)

Chapter 7

Physics of ground

The Earth's surface is a complicated affair comprising oceans, deserts, cities, forests, tundra/savannah, ice-caps, and so on. Each of these has a different albedo, rate of water evaporation/absorption, exchange of gases with the atmosphere, and so on.

7.1 Soils

Soil is composed of both rock particles and organic matter (humus) - the remains of plants and animals in various stages of decomposition. The humus serves as food for many living organisms. Within the soil is a large population of animals, plants. These break down the humus into soluble substances that can be absorbed by the roots of larger plants. Soils are categorised by the air, water, rock and humus content. The first consideration is the size of the particles that make up the soil. A set of typical sizes for various soils is given below: Much of the soil consists of air and water.

Particle Diameter (μm)	**Fraction**
> 2000	Gravel
60-2000	Sand
2-60	Silt
<2	Clay

Table 7.1: Particle Sizes in Soil

Between 30% and 70% of soil is pore space. The fraction of soil that is occupied by solids is given by:

$$\text{Fraction of soil volume occupied by solid} = \frac{\text{Mass of soil}}{\text{Mass of solids}} \times \frac{\text{Volume of solids}}{\text{Volume of soil}}$$

$$\text{Fraction of soil volume occupied by solid} = \text{bulk density} \times \frac{1}{\text{particle density}}$$

We define the porosity as the fraction of soil volume occupied by pores:

$$\text{Porosity} = 1 - \frac{\text{bulk density}}{\text{particle density}}$$

Typical bulk density and porosity can be seen in the following table:

Soil Texture	**Bulk Density (kg/m³)**	**Porosity**
Sandstone	2100	0.19
Sandy loam subsoil	1650	0.36
Sandy loam plow layer	1500	0.42
Clay loam subsoil	1450	0.44
Recently plowed clay loam	1100	0.58

Table 7.2: Bulk Density and Porosity for Different Soil Textures

7.2 Soil and hydrologic cycle

The passage of water in and through the soil is an important part of the hydrologic cycle discussed above. There are two basic problems: (i) how much water is in the soil, and (ii) how does it move through the soil. The porosity gives a measure of how much water the soil can hold. In fact, this is a gross overestimate. Water in large pores and cracks (greater than 60 μm diameter) cannot be held in the soil. This can be seen from the following argument. The flow of water through a tube depends on the tube radius, the viscosity of water and the pressure gradient trying to push the water through the tube. A simple dimensional analysis gives the Hagen-Poiseuille equation (except of course for the constant in front). This states that

$$Q = \frac{\pi r^4 dp}{8\eta dx}$$

where Q is the rate of flow of water (in m^3/s), r is the radius of the tube, η is the viscosity (units of Pa s) and dp/dx is the pressure gradient. The point is the strong dependence of the flow rate on the diameter of the tube. This badly overestimates the velocity. Pores are not straight. Water velocities through pores typical of sand grains (diameter 1000 μm) are about 10^8 times those typical of clays (pore diameter about 0.1 μm). There is an upper limit to the amount of water that a soil can hold in the long term. This is the field capacity. It is the water held in pores small enough so that friction and surface tension (see below) can resist the gravitational flow. There is also a lower limit to the amount of water that can be extracted by plant roots. If all the remaining water is held in very fine pore, the plant roots cannot extract it. This lower limit is called the Permanent Wilting Point.

7.3 Surface tension and soils

Moreover, some of the water is tightly held in the soil. This is because of the effect of surface tension. A liquid behaves as if its surface is enclosed by a skin. A molecule in the interior of a liquid experiences forces from all directions since the liquid molecules are in all directions. A molecule at the surface experiences forces only from the lower hemisphere, since above the molecule is air or vacuum (and the effect of air molecules is negligible). It therefore costs energy to make a surface (usually expressed as energy per unit area). This can also be looked at in terms of a force, the surface tension. The surface tension is the force acting across the surface pulling the molecules into the liquid. The same applies to interfaces between liquids and solids (and liquids and other liquids) since the forces between molecules in different materials are different. If the attraction between the molecules of a liquid is less than the attraction between the molecules of the liquid and that of a solid, the liquid will wet the solid. This determines the shape of the meniscus in a tube (concave or convex). Let us consider the case where the meniscus is convex (as it almost always is for water).

Let us assume that we are trying to push the water down the pore (say by forcing air into it) with a pressure p. Then the force is the pressure multiplied by the area of the tube; i.e. $\pi r^2 p$. Opposing this is the surface tension, γ. If the angle of contact between the liquid and the wall is given by θ, then the total force is $2\pi r\gamma\cos\theta$. If the system is in equilibrium, this gives the result $p = \frac{\pi\cos\theta}{r}$. It is often reasonable to set $\cos\theta$ to unity, giving

$$p = \frac{2\gamma}{r}$$

In the absence of an external pressure, p represents the tendency of water to creep along a pore from the wet to the dry end, and is often called the suction. A case of particular importance is when a column of water is supported against gravity by surface tension. In this case, it can be shown that the height of the column is given by

$$h = \frac{2\gamma}{r\rho g}$$

where ρ is the density of the liquid.

7.4 Water flow

In most cases, the two most important forces in the soil tending to move water are gravity and the changes in the suction from one area to another. It is often convenient to express these in terms of a potential. The soil-water potential, Ψ, is defined as

$$\Psi = -(depth + suction)$$

The minus sign states that the potential becomes more negative with depth or as the suction increases. This is related to the rate of water flow by Darcy's Law. The combination of gravity and surface tension sets up a potential difference in the soil. This potential difference determines the volume of water that can pass through unit cross-sectional area of soil. We write

$$\frac{d\Psi}{dx} = -\frac{QW}{\kappa}$$

where QW is the volume of water per unit cross-section and κ is the effective permeability of the soil (often known as the hydraulic conductivity). The value of κ depends on the soil type and structure (in particular on the size and distribution of pores). κ decreases rapidly as the soil gets drier. The sign ensures that water flows down the potential gradient. The balance of the effect of gravity and suction ensures that there will always be some moisture at the top of the soil where it is in contact with the air and can evaporate. Water is also returned to the atmosphere from the leaves of plants (transpiration).

The amount of water stored in the soil, the soil profile, S, is a balance between a number of factors

$$\Delta S = P - ES - T - D - R$$

where the contributions are P from the precipitation, ES from the evaporation from the soil surface, T from transpiration, D from deep drainage out of the soil layer and R from runoff. This balance is important when investigating the leaching of pollutants into the soil. The dependence of water flow rate on pore size means that the water moves faster down cracks than the surrounding small pores. This effect is hydrodynamic dispersion.

7.5 Water evaporation

Evaporation of water from the soil is an important part of the hydrologic cycle. Most of the incident solar radiation is radiated back at infrared frequencies. Of the remainder, some is conducted into the Earth, some is transferred to the overlying air and some (up to 60%) evaporates water. The principal mechanism for removing water from the land surface is turbulent transfer. It is possible to derive an expression for the evaporation rate for a simple case such as a lake surface. The Penman equation gives

$$\text{Evaporation rate} = aR + b(c_0 + c_1 v)D,$$

where R is the net radiation flux into the soil, v is the windspeed, D is the saturation deficit of the air and a, b, c_0, c_1 are disposable coefficients. Evaporation from land is much harder to describe with a simple expression. It depends strongly on the vegetation cover.

7.6 Soil temperature

The temperature range across the Earth's surface is large. The most obvious causes of variation are latitude and season. The seasonal changes are particularly big in the centres of continents. However, variations of 10°C in a single day are possible. Even slight topographic features affect the temperature. Low spots in fields have lower crop growth (by 33-50%) than other parts of the same field because frost formation is more likely. Small hills, depressions, exposure to the Sun, all affect the local temperature. The differences in air temperature just above the ground only affect the topmost layers of the soil because soil is a very poor conductor. An annual temperature variation of 30°C in air is reduced to 15°C at a depth of 1 m and 0.5°C at 8 m. Indeed, the heat conduction through the soil is so slow that the temperature cycle deep within the soil is the reverse of the seasonal variation at the surface. In the Northern hemisphere:
- At 3 m: minimum temperature is March/April, maximum temperature is September/October.
- At 7-11 m: minimum temperature is August, maximum temperature is February (but the variation is very small – see above).

Vast areas of Canada and Siberia (20% of the Earth's surface) have frozen soil. Intrusion of frost into the ground is so effective that summer heat cannot thaw the ground below 1m depth. Hence rain cannot sink into the soil and such regions become swamps in summer.

Questions:

- Discuss how characteristics of the soil pore space influence the movement of water and solutes through soil.

Chapter 8

Energy for living

In discussions regarding atmospheric physics, notable examples of environmental changes induced by human activities include the significant issue of global warming, largely driven by increased carbon dioxide emissions resulting from heightened fossil fuel combustion due to economic growth demands. Over time, the predominant energy sources have been fossil fuels—gas, oil, and coal—supplemented by a smaller proportion of wood and bio-waste. The industrial revolution has seen a thirty-fold increase in energy consumption, initially fueled by coal and later dominated by oil since 1950. Disparities in energy consumption per capita exist across different global regions.

Analysis of energy usage, particularly in industrialized nations, reveals distinct end-use patterns, with a significant portion allocated to low-temperature heating, high-temperature industrial processes, and transportation. A minor percentage is dedicated to activities reliant on electricity, such as lighting and electronic equipment. In both developed and developing nations, energy expenditure typically accounts for approximately 5

Projections suggest that continued current energy usage trends will strain oil and gas resources by the mid-21st century, leading to heightened reliance on more expensive marginal resources and increased exploration efforts. While coal reserves may last centuries, potential shortages, as exemplified by Japan and parts of Europe, indicate a growing dependence on imports from the developing world, particularly the Middle East. These developments exacerbate environmental concerns stemming from energy utilization.

The OECD identifies various stages within the fuel cycle, encompassing exploration, harvesting, processing, transport, storage, marketing, and end use, each contributing to environmental residuals, including resource consumption, effluents, physical transformations, and social/political impacts. A systems approach delineates system boundaries to assess the interactions between components and their environment, aiding in understanding the localized effects of energy systems on their surroundings.

8.1 Fossil fuels

Fossil fuels are hydrocarbon-based energy sources that are formed from the remains of ancient plants and animals over millions of years. The three main types of fossil fuels are coal, oil (petroleum), and natural gas.

The major sources of energy, including thermal power stations fueled by either fossil fuels or nuclear energy, are expected to remain dominant despite mounting concerns about global warming. Thermal power stations typically comprise a heating element, boiler, and turbine. The efficiency of a heat engine, as described by the Carnot efficiency (η), depends on the temperature difference between the hot and cold reservoirs ($\mathrm{T_h}$ and $\mathrm{T_c}$, respectively).

$$\eta = \frac{(\mathrm{T_h} - \mathrm{T_c})}{\mathrm{T_h}}, \tag{8.1}$$

In practical terms, the cold reservoir, often a river, maintains a temperature of around 15°C (288 K), while the hot reservoir can reach temperatures of $600 - 700^0$C (900 K), resulting in theoretical Carnot efficiencies of approximately 70%. However, real power stations, being irreversible Carnot engines, cannot achieve such high efficiencies, with typical coal-fired power stations reaching around 42%.

Efficiency improvements in power stations can be achieved through various means:
- Well-designed boilers can attain heat transfer efficiencies of up to 90%.
- Power stations commonly feature three steam turbines operating at different pressures, contributing to an overall efficiency of approximately 48%.
- Despite some mechanical and miscellaneous losses, these systems can still achieve considerable efficiency.

One strategy to enhance system utilization involves the implementation of combined heat and power generators (CHP). The overall efficiency of a CHP system (ηCHP) considers both the net power output and the recovered heat relative to the energy input.

$$\eta\text{CHP} = \frac{\text{netpoweroutput} + \text{heatrecovered}}{\text{energyinput}} \times 100\%. \tag{8.2}$$

CHP systems offer significant benefits, allowing for the recovery of about two-thirds of waste heat. However, their effectiveness depends on maintaining a fixed balance between heat and power, with variations in the required ratio based on practical circumstances. Implementing CHP systems often necessitates substantial investments in equipment and heat pipelines. Additionally, if integrated into district heating schemes, concerns regarding noise and pollution may arise, as the plant must be situated close to the area it serves.

8.2 Nuclear power

Nuclear power encompasses three main methods of harnessing energy from nuclear sources: thermal reactors, breeder reactors, and fusion reactors. Thermal reactors, which utilize fission of uranium or thorium isotopes, have been the primary source of nuclear power on a commercial scale. Breeder reactors, while capable of both fission and converting U238 to Pu239, remain largely experimental, as do fusion reactors, which rely on the fusion of deuterium and tritium.

All nuclear reactor designs share common features:
- Nuclear fuel, where uranium or thorium nuclei undergo fission, releasing substantial energy and surplus neutrons, leading to the potential for a chain reaction and the production of radioactive isotopes.
- Control of the chain reaction is crucial for reactor safety, achieved through the use of delayed neutrons.
- Moderators, such as graphite or water, slow down neutrons to increase the likelihood of absorption by fuel nuclei.
- Heat transfer mechanisms to extract thermal energy from the reactor core, a conventional engineering process involving heat transfer circuits and boilers.

Nuclear power generation faces significant challenges:
- Risk of major radioactive releases, exemplified by events like the Chernobyl explosion.
- Management of radioactive waste, categorized into low, medium, and high levels,

with high-level waste posing the most significant disposal challenges.
- Disposal methods include incorporating waste into glass or artificial minerals and burying them in deep repositories, although issues related to low and medium-level waste volumes also require attention.

While nuclear power offers substantial energy potential, addressing safety concerns and waste management remains critical for its sustainable utilization.

8.3 Renewable resources

The methods of harnessing energy, in the end, revolve around capturing solar energy either directly or indirectly. Tidal power is an exception as it taps into the rotational energy of the Earth, while geothermal power harnesses the internal heat of the Earth. Solar energy availability is substantial, with approximately 18,000 terawatts (TW) falling on the Earth. However, the challenge lies in efficient collection due to its low energy density. Present usage distribution is as follows:

- Hydroelectric power accounts for 6% of global energy needs.
- Biomass, including wood-burning, fulfills 1.5% of global energy requirements.
- Tidal, solar, geothermal, and wind power collectively contribute around 0.5% of global energy needs.

Thus, only hydroelectric power makes a substantial contribution to the global energy supply.

8.3.1 Hydroelectric power

The primary advantage of hydroelectric power lies in its high energy density. However, the drawbacks are significant, as hydroelectric schemes necessitate the construction of large dams, resulting in substantial social and ecological changes.

The basic principle of hydroelectric power generation is straightforward: Water passes from a dam down a tube and through a turbine, with the aim of converting the potential energy of the water into kinetic and subsequently electrical energy. The maximum power available for generation (P_0) is determined by the formula:

$$\mathrm{P}_0 = \rho \mathrm{ghQ} \tag{8.3}$$

Where: - ρ is the density of water,
- g is the acceleration due to gravity,
- h is the height drop,
- Q is the flow rate.

Alternatively, one can analyze the problem in terms of kinetic energy. If the speed of the water is v, the available power is

$$\frac{\rho \mathrm{Q} \mathrm{v}^2}{2}. \tag{8.4}$$

One common type of turbine used in hydroelectric power generation is the Pelton impulse turbine. In this system, when the water jet hits the bottom cup of the turbine, if the speed of the cup is $\mathrm{v_t}$ and the speed of the jet is v_j, the change in momentum of the fluid (and thus the force exerted on the cup) is given by:

$$\mathrm{F} = 2\rho \mathrm{Q}(\mathrm{v_j} - \mathrm{v_t}) \tag{8.5}$$

The power transferred is therefore given by:

$$\mathrm{P} = \mathrm{F}\mathrm{v_t} = 2\rho \mathrm{Q}(\mathrm{v_j} - \mathrm{v_t})\mathrm{v_t} \tag{8.6}$$

The maximum power output occurs when $\frac{v_j}{v_t} = 0.5$, resulting in the turbine being 100% efficient, with real efficiencies ranging from 50% for small units to 90% for large commercial systems.

8.3.2 Tidal power

Tidal power, akin to hydroelectric power, taps into the energy potential of tidal movements in oceans and estuaries. However, unlike hydroelectricity, tidal power is not a continuous energy source. While considerable energy is theoretically available, practical limitations include energy density—how many estuaries are suitable—and environmental concerns.

The largest tidal installation, located at La Rance in France, boasts a capacity of 240 MW. Despite the estuary's configuration being close to ideal, tidal barrages present challenges: they are expensive to build and can drastically alter the estuarine environment.

The fundamental concept involves trapping the tide behind a barrier and releasing it through turbines at low tide. If the tidal range is R and the estuary area is A, the mass of trapped water is ρAR, with its center of gravity R/2 above the low-tide level. The maximum energy per tide is thus $(\rho \mathrm{AR})\mathrm{g}(\mathrm{R}/2)$. Averaging over a tidal period τ yields a mean power available:

$$< \mathrm{P} >= \frac{\rho \mathrm{AR}^2\mathrm{g}}{2\tau} \tag{8.7}$$

However, this estimation requires further refinement to account for monthly variations (spring and neap tides). Extracting power necessitates specialized turbines designed for relatively low head. Significant power extraction close to low tide is impractical, but total output can be increased by using turbines as pumps near high tide to augment the tidal head.

8.3.3 Wind power

Wind power harnesses the kinetic energy of moving air to generate electricity, utilizing modern wind turbines typically equipped with two or three-bladed propellers, often around 33 meters in diameter. The power generated by wind turbines in a strong breeze (Beaufort scale 6) can reach 300 kW, necessitating the construction of wind farms to capitalize on wind energy resources.

An illustrative example is the Fair Isle scheme in 1982, where a 50 MW wind farm supplied 90% of the island's electricity needs. While this demonstrates the viability of wind power in certain circumstances, challenges remain, notably the mismatch between peak wind availability and peak electricity demand. Moreover, large wind farms, though effective, face opposition due to their visual impact and placement in areas of natural beauty.

The physics behind wind power is straightforward. The kinetic energy in a unit volume of air is expressed as

$$\frac{\rho \mathrm{v}^2}{2}, \tag{8.8}$$

where ρ is the air density and v is the wind velocity. The maximum power per unit area of a wind turbine is given by:

$$\frac{\mathrm{P}_0}{\mathrm{A}} = \frac{\rho \mathrm{u}^3 \cos\beta}{2} \tag{8.9}$$

where u is the wind speed and β is the angle between the wind direction and the normal of the turbine's cross-section. In principle, the maximum power available occurs when $\cos\beta = 1$ and then

$$\frac{\mathrm{P_0}}{\mathrm{A}} = \rho\frac{\mathrm{u}^3}{2} \tag{8.10}$$

The Betz limit establishes the theoretical upper bound for the coefficient of performance ($\mathrm{C_P}$), which describes the efficiency of energy extraction from the wind. The coefficient of performance ($\mathrm{C_P}$) of a wind turbine is determined by the fraction decrease in wind speed at the turbine, denoted by a, and is given by the Betz limit formula:

$$\mathrm{C_P} = 4\mathrm{a}(1-\mathrm{a})^2 \tag{8.11}$$

Where

$$\mathrm{a} = (\nu_{\mathrm{wind}} - \nu_{\mathrm{back}})/2\nu_{\mathrm{wind}} \tag{8.12}$$

is the fractional decrease in wind speed at the turbine, also known as the interference factor.

The maximum ($\mathrm{C_P}$) value occurs at $\mathrm{a} = \frac{1}{3}$, yielding a ($\mathrm{C_P}$) of 0.59. Modern wind turbines typically achieve a ($\mathrm{C_P}$) value of about 0.4, translating to approximately 95% efficiency in energy generation.

Wind power systems are particularly useful in areas where grid connectivity is costly, although they require backup storage solutions such as batteries or connection to the grid due to the variable nature of wind energy. Among renewable energy sources, wind power is closest to being competitive with conventional fossil fuels. Commercial wind projects are typically structured as wind farms, with individual turbines spaced approximately ten times the length of their blades apart, with additional buffer zones surrounding the farm.

8.3.4 Wave power

Wave power holds the potential to yield significant amounts of energy, particularly from deep water waves where the mean seabed depth (D) exceeds half the wavelength (λ). Key characteristics of such waves include:

- Surface waves manifest as irregular-phase sine waves.
- Water particles beneath the surface move in circular paths, while surface water remains on the surface.

- Amplitude of water particle motion diminishes exponentially with depth. - The amplitude of surface waves remains consistent regardless of wavelength or velocity.
- Wave breakage typically occurs when the surface slope reaches approximately 1 in 7.

The power carried by a wave derives from the alteration in potential energy as water rotates in circular paths beneath the surface. This power can be quantified as:

$$\mathrm{P} = \frac{\rho \mathrm{g}^2 \mathrm{A}^2 \mathrm{T}}{8\pi} \tag{8.13}$$

Where:
- A represents the wave amplitude at the surface,
- T denotes the wave period,
- ρ is the water density,
- g is the acceleration due to gravity.

Two primary methods proposed for harnessing wave power include the Salter duck and the oscillating column. The Salter duck comprises a cone oscillating with waves, linked to a rotary pump driving a generator. Conversely, the oscillating column employs waves to propel a trapped air column past a turbine. Despite numerous prototypes, the economic viability of wave power generation has not yet warranted full-scale commercial deployment.

8.3.5 Biomass

Biomass energy refers to the renewable energy derived from organic materials, known as biomass, that originate from plants, animals, and organic waste. This energy source is obtained through various processes such as combustion, fermentation, and chemical conversion.

Biomass, the second most significant renewable fuel source after hydroelectric power, encompasses a broad range of organic materials derived from domestic, industrial, and agricultural sources. Unlike fossil fuels, the biomass cycle is sustainable as long as each plant used as fuel is replaced by new growth. Examples of biofuels include:

Gaseous Biofuels: Utilized for heating, cooking, electricity and heat generation, and sometimes for transport. Examples include biogas (CH4 and CO2) produced from anaerobic digestion of plant and animal wastes, as well as producer gas (CO and H2) obtained from gasification of plants, wood, and wastes.

Liquid Biofuels: Primarily employed as transport fuels. Examples include oils extracted from crop seeds like rape and sunflower, esters produced from such oils, ethanol derived from fermentation and distillation, and methanol obtained from acidification and distillation of woody crops.

Solid Biofuels: Examples comprise wood from plantations, forest cuttings, timber yards, and other waste sources, charcoal generated through pyrolysis, and refuse-derived fuels such as compressed pellets.

Brazil stands as a significant user of biomass, particularly waste from the sugar-cane industry. Bagasse, the residue left after crushing the cane, and barbojo, the leaves of the cane, are extensively utilized. Roughly 67% of the 80 sugar-cane-producing countries can potentially employ these materials as fuel sources. Biomass plays a crucial role in diversifying energy sources and reducing reliance on fossil fuels.

8.3.6 Solar power

The simplest method of harnessing solar energy involves converting it into heat. A black surface exposed directly to full sunlight can absorb approximately 1 kilowatt per square meter (kW/m2).

Solar energy comes in two forms: direct and diffuse. Only direct radiation can be concentrated effectively. The amount of solar energy received at a specific location depends on factors such as latitude, time of day, and season. To maximize solar energy absorption on a surface, it should be angled so that its normal (perpendicular) points directly at the Sun. Ideally, the orientation should change throughout the day and adjust for the Sun's declination with the seasons. However, the cost of such precise adjustments is often prohibitive. A simpler approach is to position the surface to face the Sun at noon and fix the angle with the horizontal.

In some applications, concentrators can be utilized to intensify solar energy. The maximum temperature achievable with solar concentration is approximately 1150 Kelvin, although practical temperatures typically reach around 950 Kelvin. This temperature range is sufficient for efficient electricity generation and forms the basis for technologies like the solar furnace, which concentrates solar energy to achieve high temperatures for industrial processes.

8.3.7 Solar collector

A solar collector is a device designed to absorb solar radiation and convert it into heat energy, typically for heating water. Here's a breakdown of the process and calculations involved in a solar collector system:

The solar radiation is absorbed by a black surface, usually made of black chrome, with a high absorption coefficient. This surface is heated up to approximately 800°C.

The absorbed heat is then transferred to water tubes within the collector. The heated water can be used for various applications, similar to a standard boiler.

Various factors contribute to energy losses in the system, including conduction to supports, convection, and radiation. Radiation loss, governed by the Stefan-Boltzmann law, is the most significant. Glazing placed above the collector helps reduce convection losses.

Considering a modern solar collector with an area of 3 m^2 and a covering glass plate with a transmission coefficient of 90%, we can calculate the net power absorbed. The net power absorbed, P, is given by the product of the solar radiation flux (S), the area of the plate (A), and the absorption coefficient (a).

$$\mathrm{P} = \mathrm{SAta} \tag{8.14}$$

Assuming all the absorbed power is transferred to the water flowing through the collector, we can calculate the heat gained per unit time by the water using the formula:

$$\mathrm{P} = \frac{\mathrm{dQ}}{\mathrm{dt}} = \mathrm{C}\rho\mathrm{Q}(\mathrm{T_{out}} - \mathrm{T_{in}}) \tag{8.15}$$

where C is the specific heat of water, ρ is the density, Q is the flow rate, $\mathrm{T_{out}}$ is the temperature at the outlet, and $\mathrm{T_{in}}$ is the temperature at the inlet. In the second identity ρ is the density and Q the flow rate. This is, of course, an idealisation. There are bound to be some losses. These can be expressed as an effective thermal resistance of the collector. We define the resistance R as the inverse of the rate of heat transfer from the collector to the surroundings:

$$\text{Energylosses} = \frac{\mathrm{T_{out}} - \mathrm{T_{in}}}{\mathrm{R}} \tag{8.16}$$

The capture efficiency (n) of the system, representing the fraction of solar power converted into useful heat, can be expressed using the formula:

$$\text{Net Power absorbed} - \text{Energy losses} = \mathrm{n} \times \text{Incoming radiation flux} \tag{8.17}$$

This equation can be written as:

$$\mathrm{taSA} - \frac{\mathrm{T_{out}} - \mathrm{T_{in}}}{\mathrm{R}} = \mathrm{nSA} \tag{8.18}$$

Solving for n, we get:

$$\mathrm{n} = \mathrm{ta} - \frac{\mathrm{T_{out}} - \mathrm{T_{in}}}{\mathrm{RAS}} \tag{8.19}$$

It is common to define U, the energy transfer coefficient, as the inverse of the product of the resistance and the area, which gives us:

$$\mathrm{U} = \frac{1}{\mathrm{RA}} \tag{8.20}$$

Finally, we can express the capture efficiency n as:

$$\mathrm{n} = \mathrm{ta} - \mathrm{U}(\mathrm{T_p} - \mathrm{T_a}) \tag{8.21}$$

Where:
- ta is the absorption coefficient,
- S is the solar constant,
- A is the area of the collector,
- T_{out} is the temperature at the outlet,
- T_{in} is the temperature at the inlet,
- R is the thermal resistance of the collector,
- U is the energy transfer coefficient,
- T_p is the temperature of the collector,
- T_a is the temperature of the surroundings.

The capture efficiency (n) of the system is defined as the fraction of solar power converted into useful heat.

8.3.8 Solar photovoltaic

Solar photovoltaic technology converts solar radiation directly into electricity using photovoltaic cells. These cells find applications in various devices such as watches, calculators, and solar arrays for spacecraft. Practical cells are often made of amorphous silicon, with efficiencies typically ranging from 10% to 20%. This means that a panel of cells covering 1 square meter and facing full sunlight can generate around 100 to 200 watts of electricity. Therefore, a large area of solar panels is required to generate significant amounts of power.

As of 1990, the global capacity for solar photovoltaic power was approximately 50 megawatts (MW). To meet the projections set by the World Energy Council (WEC), which anticipates growing energy demands, we would need to increase this capacity by a factor of 1000 by the year 2020. This indicates a significant challenge in scaling up solar photovoltaic technology to meet the world's energy needs.

8.4 Energy demand and conservation

Energy conservation is a crucial strategy for reducing overall energy demand. While conservation efforts can vary depending on the process, addressing space heating offers a foundational example of conservation practices. To effectively approach energy conservation in this context, several factors must be considered.

Firstly, it's essential to understand the fundamental principles of heat transfer and thermal insulation. This knowledge forms the basis for designing energy-efficient systems and structures.

Additionally, energy conservation initiatives often involve trade-offs. For instance, prioritizing energy efficiency in building design may require balancing other factors such as comfort, aesthetics, and cost-effectiveness. Finding the optimal balance among these considerations is key to implementing successful conservation measures.

8.4.1 Heat transfer and thermal insulation

Heat can be transferred by: (i) conduction, (ii) convection, and (iii) radiation. Thermal insulation reduces the transfer of heat from one point to another, especially from the interior of a building to the outside. Effective insulation reduces the amount of heat that has to be supplied.

The effectiveness of an insulator is measured by its thermal conductivity, κ. This is defined from the Fourier heat equation. Fourier asserted that heat flow per unit cross-sectional area, J, is proportional to the temperature gradient, i.e.,

$$J = -\kappa \frac{dT}{dx}$$

The negative sign states that heat flows down the temperature gradient: from hot to cold. Typical values of κ ($\mathrm{Wm^{-1}K^{-1}}$) are Al (160), steel (50), brick (0.84). Air is a good insulator but should not be in motion or convection will transfer heat.

Heat transfer by convection is usually divided into: (i) natural convection (where the fluid moves without any forcing), and (ii) forced convection (where the fluid is moved by draughts). Natural convection is described by Newton's law of cooling:

$$J = k(T - T_0)$$

where T is the temperature of the object, T_0 is the ambient temperature, and k is the convection coefficient.

Heat loss by radiation is given by the Stefan-Boltzmann law.

Since in all cases, we are interested in the problem of heat transfer through the parts of the building, it is convenient to choose a measure that does not depend on the mechanism. Engineers use a measurement to quantify the thermal behaviour of a structural element. The U-value (or thermal transmittance coefficient) is the rate at which heat flows through an area of 1 m^2 of an element when the temperature change across it is $1°C$. Clearly, this is most easily related to the thermal conductivity. We can express the heat transfer equation as a finite difference equation,

$$J = -\kappa \frac{\Delta T}{\Delta x}$$

Thus, from the definition of U given above,

$$U = -\frac{J}{\kappa} = \frac{\Delta T}{\Delta x}$$

Radiation losses are forced into this form. For convection (as treated above) $U = k$.

The total U-value for a complex system is obtained by using Kirchoff's law to sum the resistances. We define the thermal resistance $R = \kappa^{-1}\Delta x$. There are also resistances due to the presence of interfaces. These are given by heat transfer coefficients, h. Thus, the total U-value for a complex wall with heat transfer coefficients h_{in} and h_{out} for transfer into and out of the wall respectively is given by:

$$\frac{1}{U} = \frac{1}{h_{\text{in}}} + \sum_j R_j + \frac{1}{h_{\text{out}}}$$

For example, a single window has $U = 5.7\,\text{Wm}^2\text{K}^{-1}$ Since a double-glazed window has a 2 cm air space, U is (roughly) halved. The lower the value of U, the better the insulation.

Example: A cavity wall.

Material	Thickness (m)
Brick	0.1
Air	0.1
Concrete	0.1
Plaster	0.1

Let us assume that the thermal conductivities are:

$$\begin{aligned}
\text{Brick} &: 0.8\,\text{Wm}^{-1}\text{K}^{-1}, \\
\text{Concrete} &: 0.2\,\text{Wm}^{-1}\text{K}^{-1}, \\
\text{Plaster} &: 0.17\,\text{Wm}^{-1}\text{K}^{-1},
\end{aligned}$$

and that all the materials are 0.1 m thick.

The thermal resistances are: plaster/air interface: $0.12\,\text{m}^2\text{KW}^{-1}$, brick/air interface: $0.16\,\text{m}^2\text{KW}^{-1}$, cavity (including interfaces): $0.19\,\text{m}^2\text{KW}^{-1}$. The thermal resistances of the materials are: brick $= 0.1/0.8 = 0.125$, concrete $= 0.1/0.2 = 0.5$, plaster $= 0.1/0.17 = 0.59$. Hence, the U value is given by

$$U = \frac{1}{(0.125 + 0.5 + 0.59 + 0.19 + 0.06 + 0.12)} = 0.63$$

ignoring the two interfacial transfer coefficients. Typical values for these would be $h_{\text{in}} = 7\,\text{Wm}^{-2}\text{K}^{-1}$ and $h_{\text{out}} = 18\,\text{Wm}^{-2}\text{K}^{-1}$. This gives a final U-value of $0.56\,\text{Wm}^{-2}\text{K}^{-1}$.

8.4.2 Heat loss in buildings

Heat loss in buildings is influenced by various factors, including insulation, the surface area of external building components, temperature differentials between indoor and outdoor environments, ventilation rates, and exposure to climatic conditions such as wind. These factors can be quantified using U-values, which measure the thermal conductivity of building materials.

There are two main types of heat loss in buildings: fabric loss and ventilation loss.

1. Fabric loss refers to heat lost through the external structure of the building, including walls, floors, ceilings, and windows. It can be quantified using the equation:

$$P = UA(T_{\text{in}} - T_{\text{air}})$$

where:
- P is the power loss,
- U is the effective U-value for the building,
- A is the surface area of the building component,
- $T_{\text{in}} - T_{\text{air}}$ is the temperature difference across the building component.

Windows typically have the highest U-values, but walls contribute the most to overall heat loss due to their larger surface area.

2. Ventilation loss occurs as a result of air exchange between the interior and exterior of the building. The rate of heat loss through ventilation can be expressed as:

$$Q = mC_p\Delta T$$

where:
- m is the mass of air,
- C_p is the specific heat at constant pressure,
- ΔT is the temperature difference between indoor and outdoor air.

For a room with volume V and air replacement time t, the rate of heat loss due to ventilation is:

$$\frac{V\rho C_p \Delta T}{t}$$

where ρ is the density of air.

In addition to fabric and ventilation losses, buildings experience heat gain from various sources:
- Solar heat through windows, walls, and roofs
- Body heat from occupants
- Heat generated by lighting, electrical appliances, and cooking
- Heat from water heating systems

When designing heating systems for buildings, it's crucial to account for all these factors to ensure energy efficiency and comfort.

Questions:

• The solar constant has a value which varies from 1420 W/m^2 in December to 1330 W/m^2 in June. Suggest why the solar constant is different in December from June.