

# Chapter 2

## First Principles of Meteorology

### Contents

2.1	General Aspects of Meteorology .....	68
2.2	Vertical Structure of the Temperature and Conditions of Atmospheric Stability .....	70
2.2.1	Dry Vertical Temperature Lapse Rate .....	71
2.2.2	Wet Vertical Temperature Lapse Rate .....	74
2.2.3	Temperature Inversion .....	75
2.3	Atmospheric Variability – Air Masses – Fronts .....	77
2.3.1	Air Masses .....	77
2.3.2	Classification of Air Masses .....	78
2.3.3	Fronts .....	79
2.3.4	Wave Cyclone .....	85
2.4	Turbulence – Equations for the Mean Values .....	86
2.5	Statistical Properties of Turbulence .....	87
2.6	Atmospheric Temperature .....	92
2.6.1	Temperature Season Variability .....	92
2.6.2	Temperature Daily Variability .....	96
2.6.3	Heating of the Earth's Surface and Heat Conduction .....	98
2.6.4	Distribution of Temperature in the Air .....	100
2.7	Humidity in the Atmosphere .....	101
2.7.1	Mathematical Expressions of Humidity in the Atmosphere .....	102
2.7.2	Dew Point .....	104
2.7.3	Clouds in the Atmosphere .....	106
2.7.4	Precipitation .....	107
2.7.5	Study of Precipitation Scavenging .....	109
2.8	Applications and Examples .....	112
	References .....	117

**Abstract** The second chapter examines general aspects of meteorology including conditions of atmospheric stability in conjunction with the vertical temperature lapse rate. It also studies large-scale weather changes which are related to changes in pressure systems. Air masses have different thermodynamic characteristics based on their origin and the morphology of the surfaces above which they move. A classification of the air masses is created, together with a mathematical description of the statistical properties of the air masses' transport. Furthermore, the temperature and humidity variability in the atmosphere is studied. Finally the clouds in the atmosphere and general aspects of precipitation are also studied.

2.1 General Aspects of Meteorology

The study of atmosphere dynamics is subsumed under the science of meteorology. The atmosphere contains thousands of chemical species in trace quantities (ppm to ppt levels) (Finlayson Pitts and Pitts 1986; 2000). Therefore the troposphere can be viewed as a huge container that includes gaseous and particulate matter pollutants. The atmosphere is a dynamic system with continuous exchange of its gaseous components between the atmosphere and the earth’s surface, including the vegetation and the oceans. Emissions of pollutants are transported into the atmosphere at long distances from their sources. The dynamic of the atmosphere and the chemical reactivity of the pollutants, as well as the size of particulate matter, determine their residence time and their effects on humans and ecosystems (Seinfeld and Pandis 2006). Table 2.1 presents the different spatial scales of pollutant transport in the atmosphere and related physico-chemical processes.

Figure 2.1 presents the different time and length scales related to atmospheric processes ranging from molecular diffusion to climatic impacts. In the atmosphere the chemical composition of atmospheric species can be divided into four main groups, namely sulfur, nitrogen, carbon and halogen containing compounds (Finlayson-Pitts and Pitts 1986; Seinfeld and Pandis 2006). Of course there are chemical compounds in the above groups that include atoms from other groups such as compounds that include both sulphur and carbon atoms. The chemical compounds which are emitted into the atmosphere eventually are removed and there exists a cycle for these compounds. This cycle is called biogeochemical cycle of the compound. The term “air pollution” is used when chemical compounds emitted from mainly anthropogenic activities are at concentrations above their normal ambient levels and have measurable effects on humans and ecosystems. In addition, Fig. 2.1 shows the time and spatial scales related to air pollution, turbulence, clouds, weather and climate.

Understanding of the complex sequence of events starting from the emissions of air pollutants to the atmosphere with human health effects as a final event is necessary for the prognosis of potential risk to humans from specific chemical compounds and mixtures of them (see Fig. 2.2). Furthermore, understanding of the chemical composition/size distribution characteristics of particulate matter (PM) and the chemical reactivity of gaseous pollutants together with their indoor-outdoor

**Table 2.1** Spatial scales of pollutant transport in the atmosphere and related phenomena

Scale	Dimension	Examples of physical and chemical processes in the atmosphere
Molecular scale	<< 2 mm	Molecular diffusion
Microscale	2 mm–2 km	Industry emissions, clouds
Mesoscale	2–2,000 km	Cloud coalesce, storms, air pollution at urban centers
Synoptic scale	500–10,000 km	Low and high pressure systems, ozone hole in the Antarctica
Global scale	> 10,000 km	Decrease of stratospheric ozone, planetary wind systems

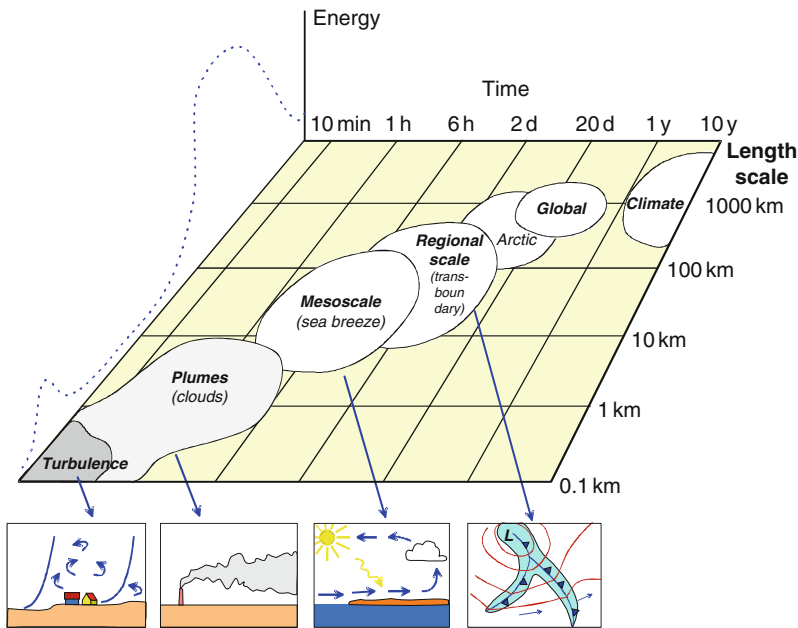


Fig. 2.1 Time (s), energy and length (m) scales in the atmosphere and related phenomena

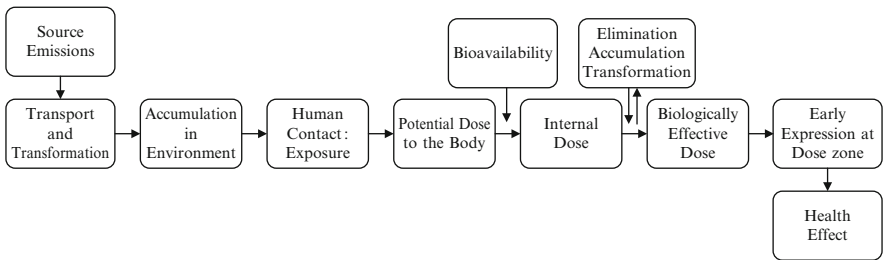


Fig. 2.2 Schematic representation of the complex sequence from emissions of air pollutants to health effects (Adapted from Liroy 1990)

characteristics and their relation to human exposure and internal dose are necessary steps for the quantification of human exposure to air pollutants.

Many problems of air pollution occur on several scales, such as the acidification problem which extends from the mesoscale to a regional scale. The majority of air pollution episodes occur in the lower part of the atmosphere, which is called the planetary boundary layer as discussed in Chapter 1. This layer is defined as the lowest part of the troposphere which is affected from the surface forces in time

scales of 1 h or lower. The structure of the boundary layer is not static but dynamic and contains usually the first 1,000 m from the Earth's surface.

## 2.2 Vertical Structure of the Temperature and Conditions of Atmospheric Stability

Meteorology examines the movement of air molecules in the atmosphere. Atmospheric stability is examined under the conditions in which there is a small displacement of an air volume with the application of an external force. In the case that the external force results in bringing the air volume to its original position prior to its displacement, then there are stable conditions in the atmosphere. On the contrary, if the external force brings the air volume away from the original position, there are unstable conditions.

Stability in the atmosphere is dependent on the vertical profile of temperature and humidity of ambient air. Warm air has lower density than cold air and therefore it is lighter. A similar situation occurs for humid air which has lower density than dry air and therefore is lighter. Consequently, a warmer or more humid air volume than the surrounding ambient air is characterized as unstable and will ascend into the atmosphere. On the contrary, an air volume that is colder or drier than the surrounding ambient air is characterized as stable and will descend into the atmosphere until it reaches equilibrium.

The stability conditions in the atmosphere are related to the atmosphere's ability to mix and spread out pollutants. These conditions determine also the turbulent conditions in the atmosphere and the cloud formation.

Atmospheric air absorbs less heat than the Earth's surface due its lower heat capacity. Therefore the layer of the atmosphere which is closer to the Earth's surface receives more energy, and consequently more heat, than the above layers. Due to the heating these layers of air become lighter than the above layers and are lofted above with consequent expansion and cooling. Since this air volume expansion occurs at small time intervals without significant heat exchange with the surrounding air environment, it is an adiabatic process. Of course the process is not pure adiabatic since the air masses are not thermally insulated. However, since the expansion of the air masses happens quickly and the heat exchange with conduction and radiation is slow, the process of adiabatic expansion in the atmosphere is assumed. Therefore, when an air mass is moved to an area with lower pressure it is adiabatically expanded and cooled. On the contrary, when an air mass is moved to an area with a higher atmospheric pressure it is contracted adiabatically and heated. Indeed, observations have shown the importance of adiabatic processes in the atmosphere in relation to weather.

In the International Standard Atmosphere (ISA) the temperature is decreased with height at a rate of  $0.65^{\circ}\text{C}/100\text{ m}$ . This rate is called the temperature lapse rate. The determination of the temperature lapse rate is presented in the following subsections.

### 2.2.1 Dry Vertical Temperature Lapse Rate

The calculation of the temperature lapse rate for a volume of dry air is examined here assuming adiabatic expansion as discussed earlier. The first law of thermodynamics can be expressed as:

$$dU = dQ + dW, \quad (2.1)$$

where, U, Q and W are the internal energy of the air volume, its heat and produced work respectively. The internal energy is given from the expression

$$dU = C_v dT, \quad (2.2)$$

where,  $C_v$  is the thermal capacity of the system at constant volume. The expression which gives the work is given by

$$dW = -p dV. \quad (2.3)$$

The law of ideal gasses can be also expressed as

$$pV = m R T / M_{air} \Rightarrow d(pV) = m R dT / M_{air} = p dV + V dp \quad (2.4)$$

where m is the air mass.

Since adiabatic conditions occur:  $dQ = 0$ .

Combining equations (2.1) to (2.4) it can be found that

$$C_v dT = V dp - \frac{m R dT}{M_{air}} = \frac{m R T}{M_{air}} \frac{dp}{p} - \frac{m R dT}{M_{air}} \Rightarrow \frac{dT}{dp} = \frac{m R T / M_{air} p}{C_v + m R / M_{air}}. \quad (2.5)$$

In Chapter 1 the change of pressure versus height was described with the hydrostatic equation:  $\frac{dp}{dz} = -\frac{M_{air} g p}{R T}$ . With the combination of the hydrostatic equation and equation (2.5) it can be concluded that:

$$\frac{dT}{dz} = -\frac{m g}{C_v + m R / M_{air}} = -\frac{g}{\hat{C}_v + R / M_{air}}, \quad (2.6)$$

where  $\hat{C}_v$  is the thermal capacity at constant volume per unit mass. Finally, it can be written that  $\frac{dT}{dz} = -\frac{g}{\hat{C}_p}$  since  $\hat{C}_p - \hat{C}_v = \frac{R}{M_{air}}$ .

The ratio  $\frac{g}{\hat{C}_p}$  for dry air is equal to 0.976°C/100 m, has the symbol  $\Gamma$  and is called dry lapse rate.

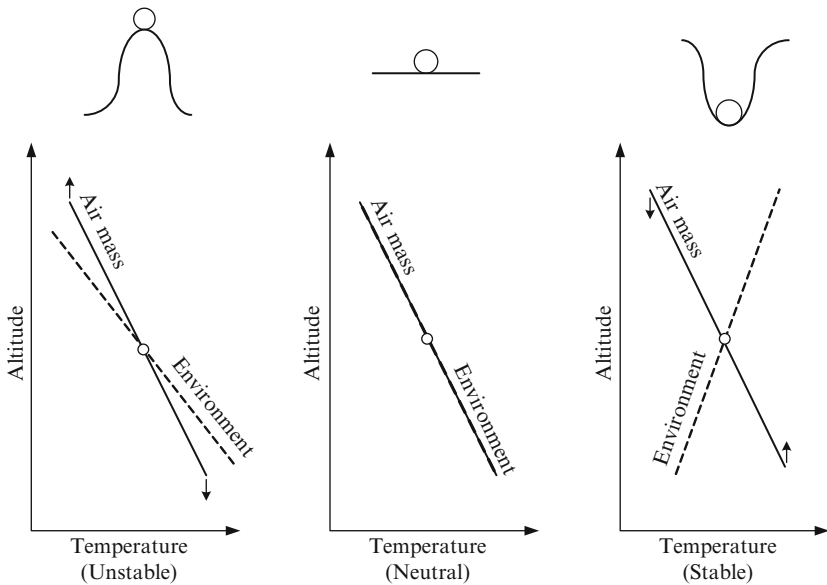
If  $\Lambda$  is the dominant lapse rate in the atmosphere, then the following cases of stability exist for the volume of air:

$$\Lambda = \Gamma, \text{ neutral}$$
$$\Lambda > \Gamma, \text{ unstable}$$
$$\Lambda < \Gamma, \text{ stable}$$

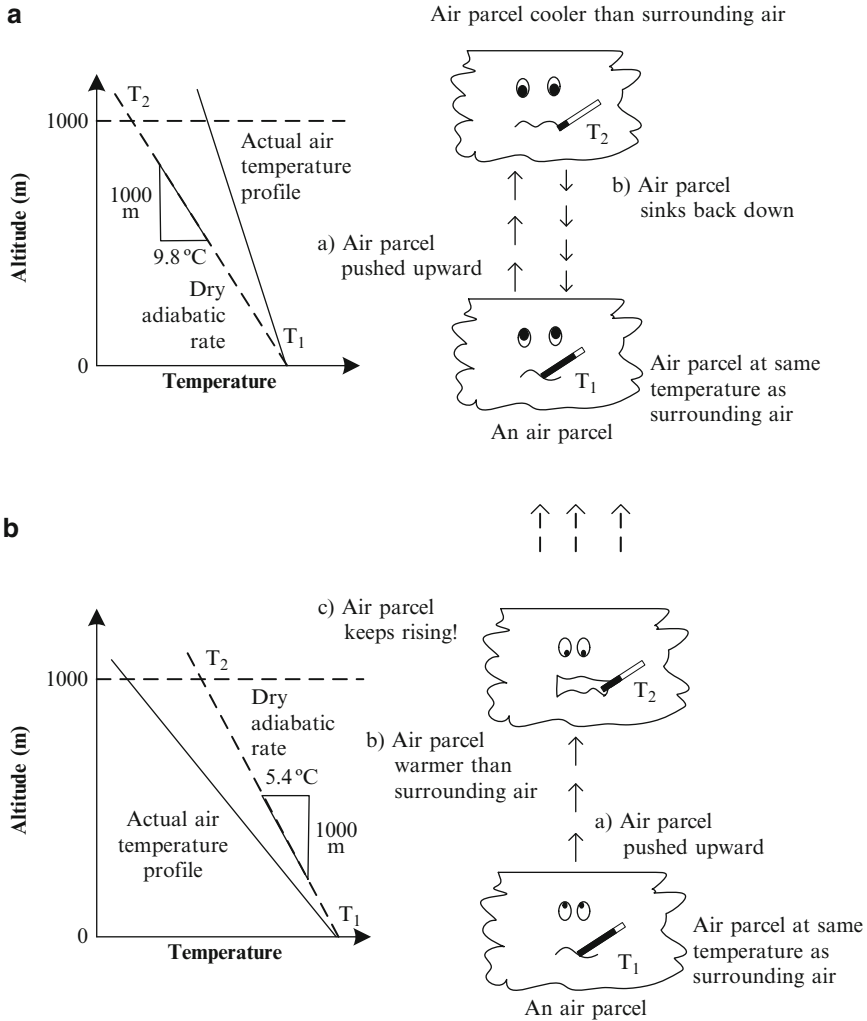
Usually, unstable conditions exist in the first 100 m from the Earth’s surface on a sunny day. Neutral conditions exist at day or night with clouds and wind and stable conditions near the Earth’s surface at night.

Figure 2.3 shows examples of different stability conditions in the atmosphere. The solid and semi-continuous lines refer to the lapse rates of the air volume under study and the atmosphere respectively. The circle denotes the air volume under study and its position in the atmosphere. At this position the air volume temperature is equal to the atmosphere’s temperature. Unstable conditions occur when the temperature of the rising air volume is higher than the temperature of the surrounding air. In this case the air volume will accelerate upwards. Stable conditions occur when the temperature of the rising air volume is lower than the temperature of the surrounding air. In this case the air volume tends to return to its initial position of equilibrium. Neutral conditions prevail when the temperature of the rising air volume is the same as the temperature of the surrounding air. In this case the air volume follows the movement of the surrounding atmosphere.

At the upper part of the Fig. 2.3 is shown a relative gravitational example for the stability conditions where a ball is located at the top of a hill (unstable condition), at



**Fig. 2.3** Graphical representation of the relation between the temperature and the height for an air volume for unstable, neutral and stable conditions in relation to the surrounding atmosphere (environment) (Adapted from Hanna et al. 1981)



**Fig. 2.4** The relation of the temperature versus height for an air volume for (a) stable and (b) unstable conditions (Adapted from Hemond and Fechner-Levy 2000)

a flat surface (neutral condition) and at the base of a valley (stable condition). Another representation of the unstable and stable conditions in the atmosphere is shown in Fig. 2.4.

A more widely accepted methodology for the calculation of atmospheric stability has been introduced by Pasquill. The methodology is based on measurements of the wind speed at a height of 10 m and intensity of the Sun's radiation during the day and cloud cover during the night (Table 2.2). However, a more practical approach is the use of a radiosonde for determination of the vertical profile of several meteorological parameters such as temperature and pressure.

**Table 2.2** Stability conditions in the atmosphere using the Pasquill methodology

Wind velocity at height of 10 m (m/sec)	Day			Night	
	Incoming sun's radiation intensity			Cloud cover	
	High	Medium	Low	>4/8	<3/8
<2	A	A-B	B	-	-
2-3	A-B	B	C	E	F
3-5	B	B-C	C	D	E
5-6	C	C-D	D	D	D
>6	C	D	D	D	D

A Very unstable

B Moderate unstable

C Slightly unstable

D Neutral (it is applied for conditions of total cloud cover both day and night)

E Slightly stable

F Moderate stable

**Table 2.3** Equivalence of stability classes with the methodologies of the vertical temperature change and the Pasquill methodology

Stability class	Vertical temperature change dT (°C/100 m)	Pasquill class
Unstable	dT < -1	A + B + C
Neutral	-1 ≤ dT < 0	D
Slightly Stable	0 ≤ dT < 1	E
Stable	dT ≤ 1	F

Another methodology for determination of stability classes is based on examination of the vertical temperature profile. This methodology is used widely for the application of Gaussian models and the equivalence of this stability class methodology with the Pasquill methodology is presented in Table 2.3.

### 2.2.2 Wet Vertical Temperature Lapse Rate

When the air contains water vapor, the thermal capacity  $\hat{c}_p$  of air has to be corrected. If  $w_v$  is the ratio of the mass of water vapor to the mass of dry air in a specific air volume, then the new thermal capacity coefficient  $\hat{c}_p'$  is given by the expression:

$$\hat{c}_p' = (1 - w_v)\hat{c}_{pa} + w_v\hat{c}_{pv} \tag{2.7}$$

where,  $\hat{c}_{pv} > \hat{c}_{pa}$  and therefore  $\hat{c}_p' > \hat{c}_{pa}$ . The symbol  $a$  refers to air and  $v$  to water vapor.

Therefore the rate of cooling of rising humid air inside a cloud is smaller than that of dry air. The humid air volume will continue to rise until the partial pressure of the water vapor becomes equal to the equilibrium water vapor pressure.

This condition will lead to the condensation of water vapor. If  $\Delta H_v$  is the heat of sublimation, then the release of heating due to the condensation of water vapor is given by the expression

$$dQ = -\Delta H_v m dw_v. \quad (2.8)$$

Following the same derivation as in the case of dry air, using the first law of thermodynamics, it can be concluded that

$$\begin{aligned} C_v dT &= -\Delta H_v m dw_v + \frac{m R T}{M_{air}} \frac{dp}{p} - \frac{m R dT}{M_{air}} \\ \Rightarrow \left( C_v + \frac{mR}{M_{air}} \right) dT &= -\Delta H_v m dw_v + \frac{m R T}{M_{air}} \frac{dp}{p} \end{aligned}$$

with a final result being

$$\frac{dT}{dz} = -\frac{g}{\hat{c}_p} - \frac{\Delta H_v}{\hat{c}_p} \frac{dw_v}{dz}. \quad (2.9)$$

The term  $\frac{dw_v}{dz}$  has negative value for a rising volume of air where water vapor condensation takes place inside. Therefore the cooling rate of a humid air inside clouds is lower than that of dry air. The reason is that with the increase of height the percentage of water vapor is decreasing due to its condensation. The term  $\frac{dw_v}{dz}$  is dependent on temperature since the equilibrium vapor pressure of water vapor is increasing considerably with the temperature.

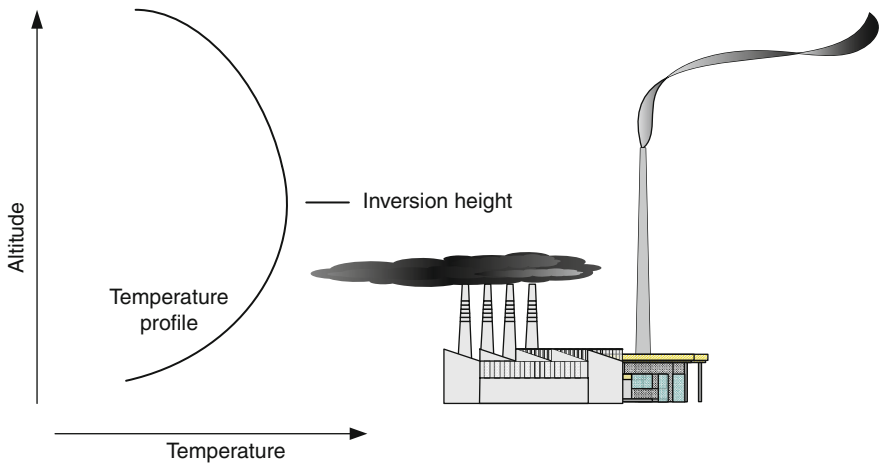
### 2.2.3 Temperature Inversion

As we have examined in Chapter 1, on many occasions the temperature of air instead of decreasing with height increases at some places in the atmosphere. This is called temperature inversion. The atmosphere is in stable equilibrium inside the inversion layer and therefore it does not favor vertical movements of air. The temperature inversion may result in increased concentration of air pollutants at the lower layers of the atmosphere.

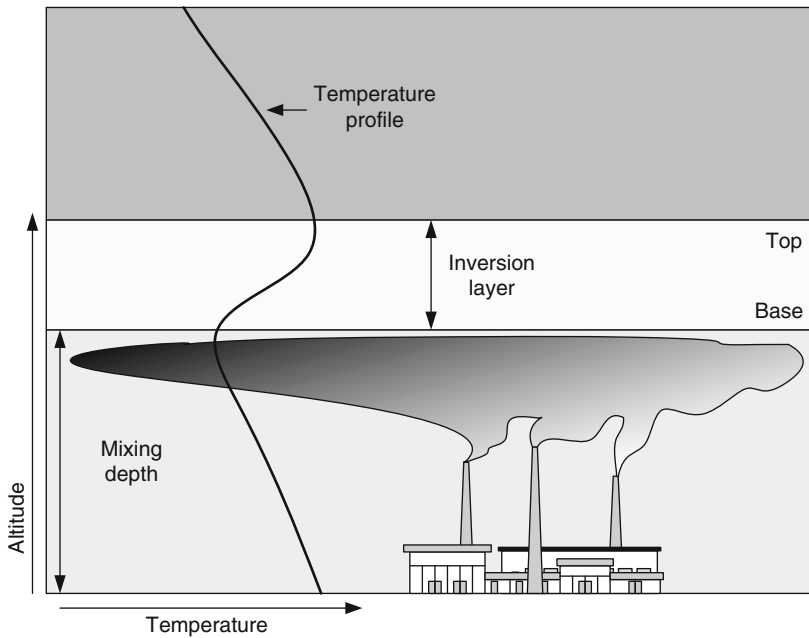
There are three main factors which may result in the occurrence of thermal inversions:

- The cooling of lower atmospheric layers (radiation or surface inversion).
- The adiabatic warming of descending air (subsidence inversion).
- Horizontal transport of warm or cold air.

A typical profile of a plume during a subsidence inversion is shown in Fig. 2.5 and in the case of radiation inversion in Fig. 2.6. The layer which extends from the



**Fig. 2.5** The plume from the lower chimneys is trapped inside the inversion, whereas, the plume from the tall chimney which is located above the inversion is rising, mixing and transported



**Fig. 2.6** The inversion height is not allowing the escape of gaseous pollutants under it. If the inversion height is becoming lower, then the mixing height is reduced and the pollutants are trapped in a smaller air volume

Earth's surface up to the base of the inversion layer is called the mixing layer. The height of this layer is called the mixing height.

It can be noted that the relatively unstable air below the inversion allows vertical mixing of air pollutants up to the base of the inversion layer. The stable air in the inversion layer does not allow vertical mixing inside the inversion layer and acts as an obstacle to the pollutant entrainment upwards. If the inversion moves upwards, then the height of the mixing layer is increasing and the pollutants will mix in a bigger air volume and, on the other hand, if the inversion moves downwards then the mixing height will be lower and the pollutants will be concentrated in a smaller air volume. In the latter case the concentration of pollutants is increasing and in urban areas may lead to concentrations above the health limits. Since the atmosphere tends to be usually more unstable during the afternoon and more stable in the morning, then there is higher mixing height at the afternoon and lower early in the morning.

## **2.3 Atmospheric Variability – Air Masses – Fronts**

Weather changes of large scale are related to changes of the pressure systems. The first step for weather prognosis is given by the Scandinavian School of meteorology which realized the importance of the formation and transport of High and Low pressure systems. The next step was the discovery of the characteristics of the air masses and front zones, the areas in which air masses are met with different thermodynamic characteristics. This led to a more detailed study of the air masses which form the basis for the weather phenomena.

Generally the meteorological conditions which are persistent in an area at specific time are dependent on the characteristics of the air masses which pass above this region or from the interactions of two different air masses which meet there. Therefore the weather above a region is quite uniform with small differences which are due to local morphological characteristics of the region such as the topography, the vegetation and distribution of land and sea.

### **2.3.1 *Air Masses***

Air masses are large bodies of atmospheric air which have similar properties of temperature and humidity distribution. It is possible for the diameter of an air mass to be larger than 1,500 km, covering large continental and ocean regions. Its height may reach the tropopause. With the air masses is accomplished the general atmospheric circulation and the transport of large quantities of heat from the equator to the poles. Inside an air mass, important factors are its humidity content, its temperature and especially the temperature change with height. The quantity of humidity influences the cloud type and consequently the rain quantity. The vertical temperature distribution affects the stability of the air mass.

In order for an air mass to attain similar temperature and humidity characteristics, it is necessary to remain for several days stagnant above a region which also has similar characteristics. This region is called the source of the air mass. At which degree the air mass attains the characteristics of the source region is dependent on its residence time and the temperature difference between the initial air temperature and the temperature of the region's surface. The air mass during its residence above the region becomes at a certain degree homogeneous. Under the influence of wind flow the air mass starts to move and retains during its transport at high degree the characteristics of its regional source.

From a thermodynamic point of view the air masses are classified in two categories, the warm and cold ones. The warm air masses are warmer than the surface above which they are transported and for this reason are becoming colder at their base and are also stable. The freezing of air masses is more effective on the layers which are contiguous to the surface and extends slowly above, mainly due to turbulent movements and not due to heat transfer. This is also the reason for the formation of temperature inversions. When the winds are weak there are often formations of mist and dew, whereas, with stronger winds there is the formation of low clouds (Stratus), below the upper limit of the temperature inversion.

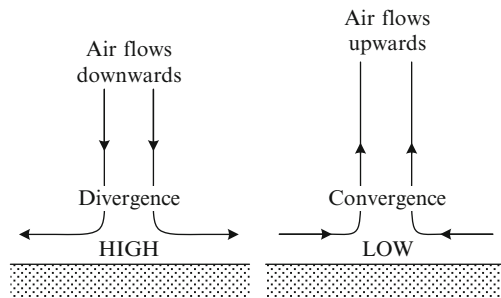
Cold air masses are colder than the Earth's surface and for this reason receive heat from below. The heating of those layers which are located close to the surface results in a sudden drop of their temperature versus height. This temperature drop results also in an increased temperature lapse rate and therefore unstable conditions and enhanced ascending air movements. If these cold air masses contain sufficient quantities of humidity or if they are supplied with humidity from the warmer layers below, then there is a formation of clouds of vertical development which also results in intense rainfall.

### ***2.3.2 Classification of Air Masses***

The classification of air masses is based on the following criteria:

- The source of the air mass. Air transported above oceans absorbs humidity and tends to be saturated in the lower layers. On the contrary, continental air masses remain dry since there is not enough water quantity for evaporation on the surface.
- The trajectory which is followed above the Earth's surface. The polar air which is transported at lower latitudes is receiving heat from below and becomes unstable. On the contrary, the tropical air which is transported at higher latitudes becomes more stable since it becomes colder at its base.
- If the air is characterized as diverging or converging. An air mass which is affected from the divergence of air at a high pressure system at the Earth's surface will move downwards at a low rate and converted to a warmer, drier and

**Fig. 2.7** The descending air, which is resulted from divergence at the surface is stable, whereas the ascending air which is resulted from convergence at the surface is unstable



more stable air mass. On the contrary, an air mass which is affected from the convergence of air at a low pressure system at the Earth's surface will move upwards and will be converted to a colder and more unstable air mass (Fig. 2.7).

Based on the above characteristics one can distinguish two main categories of air:

- Polar air masses at high geographical latitudes and
- Tropical air masses at low geographical latitudes.

These two basic categories are further divided into continental and ocean masses based on whether the air mass moves above a continent or a sea as follows:

- Polar Maritime air (Pm)
- Polar Continental air (Pc)
- Tropical Maritime air (Tm)
- Tropical Continental air (Tc)

In addition to the above basic air masses, there are also the Arctic masses. Sources of the arctic masses are the arctic regions close to the poles. These are cold air masses through all seasons and especially during winter. The behavior and alteration of the arctic air masses that intrude at lower latitudes is dependent on the part of the Earth's surface above which they are transported. In general if they are transported above oceans, then the lower layers become warmer and receive large quantities of water vapour. This results in the formation of unstable clouds of vertical development and extension of the weather characteristics in larger areas. Transport above cold continental areas results in stable air masses with mild weather characteristics, the main feature being intense cold.

### 2.3.3 Fronts

We pointed out previously that air masses have different thermodynamic characteristics based on their source of origin and the morphology of the surfaces above which they move. When two air masses with different characteristics come in

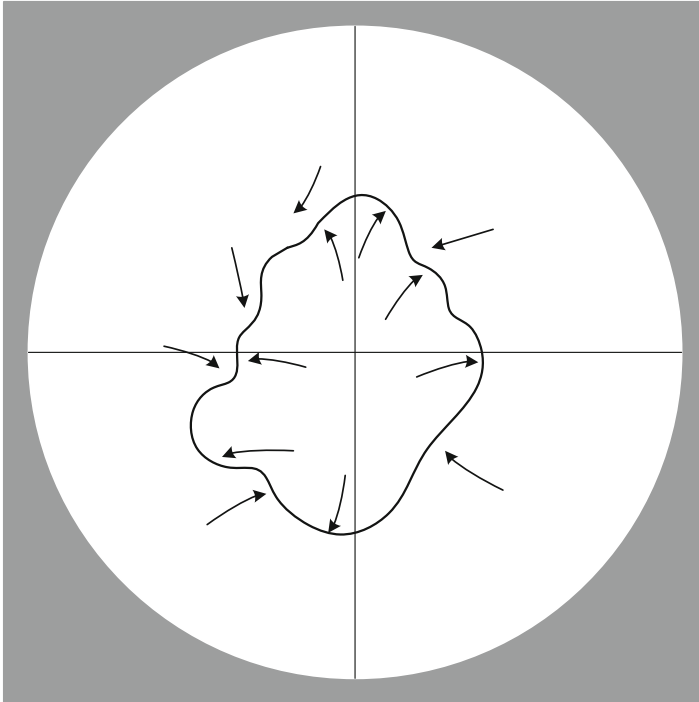
contact, they mix very slowly and form an inhomogeneous surface whose vertical profile is called a frontal surface. The cross section of this surface with the Earth's surface is called a front.

The frontal surfaces have small depth and are steep in relation to the horizon, with the warmer mass always being above the colder one. The most important fronts are the following:

### 2.3.3.1 Polar Front

At the mid-latitudes, polar cold air masses collide with tropical warm ones. The separation zone between these tropical and polar air masses is called a polar front. The position of the polar front is variable. The strong polar air moves the warm tropical air at some places whereas, in some other places the polar front declines at northern positions due to pressure from the tropical air. As a result the polar front has a corrugated form as shown in Fig. 2.8.

However, the polar front seldom is found as a continuous zone encompassing the whole hemisphere as shown in Fig. 2.8. There are locations around the hemisphere where the transition between the polar and tropical air masses is so smooth that the dividing curve does not appear. Therefore the polar front is not continuous,



**Fig. 2.8** Semi-hemispheric view of the polar front

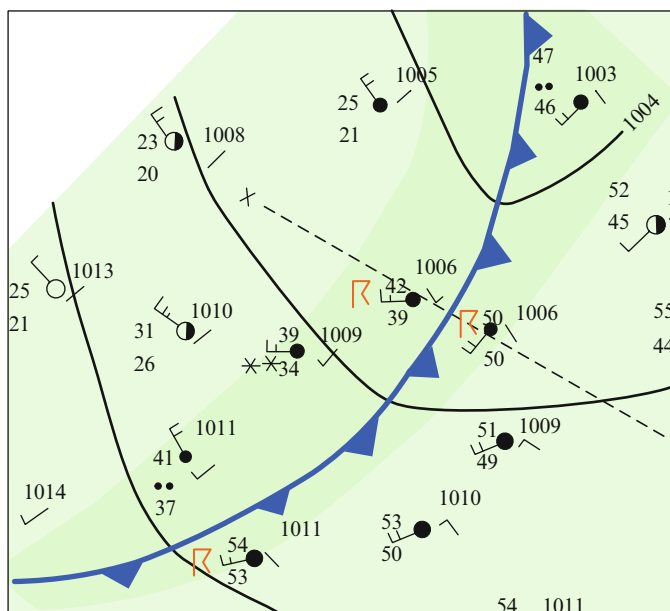
especially during the summer period in the northern hemisphere where the front is pressed northerly of the 60° parallel. During winter the polar front that is located usually at medium latitudes moves south and invades the tropical zones.

### 2.3.3.2 Cold Front

Even though the polar front represents a main zone of discontinuity in each hemisphere, fronts can be formed at any location on Earth if there is a coalescence of air masses with different thermodynamic characteristics. During the meeting of the air masses, two types of fronts are formed, the warm and cold fronts.

When two air masses (a cold and a warm one) come in contact and move so that the cold mass displaces the warm mass, then the surface that divides them is called a cold frontal surface. A side view of a cold front is shown in Fig. 2.9.

In a cold front, the warm mass which is located at the lower layers moves slower than the cold air. Therefore the cold air which moves quicker and penetrates through the lower levels of the warm air, causes a violent vertical upward movement. It has to be noted that the warm air is ascending at both warm and cold fronts and is responsible for the weather changes. It is furthermore interesting to refer to the characteristics of the warm and cold fronts, since their dynamics determine the extent of cloud coverage, the intensity and duration of rain and finally the direction and velocity of the wind.



**Fig. 2.9** Surface weather associated with a cold front. The dark band denotes the region of the weather phenomena

The peak of a cold frontal surface in relation to the Earth’s surface is close to 1/50 and the frontal weather covers a narrow region ranging from 30 to 50 miles. The physical process which occurs for the formation of the cold front is related with the movement of warm air which is transported faster than the warm air and, because it is heavier, is located under it. Therefore the warm air which is lighter has an upward movement along the frontal surface. During the upward movement process it is cooled adiabatically due to expansion and reaches saturation. Furthermore there is a condensation process and cloud formation. Since the ascending warm mass is transported quickly, there is formation not only of stratocumulus clouds but also of clouds of vertical development (storms). The weather phenomena which are related with a cold front are intense (intense rain, storms, hail) and of small time scale.

The distribution of clouds as well rain is dependent on the atmosphere’s stability and the humidity percentage of the rising warm air. The limit between the two air masses on the Earth’s surface is depicted in meteorological maps as a line with a blue shading and solid triangles which point towards the moving front. The cold front moves quite quickly.

2.3.3.3 Warm Front

When two air masses (warm and cold) are in contact and moving so that the warm mass forces the cold mass to move, then the surface which divides them is called a warm frontal surface. The intersection of this surface with the Earth’s surface is called a warm front (Fig. 2.10). Since the warm air mass is lighter, it ascends above the cold air and cools adiabatically resulting in extended condensation.

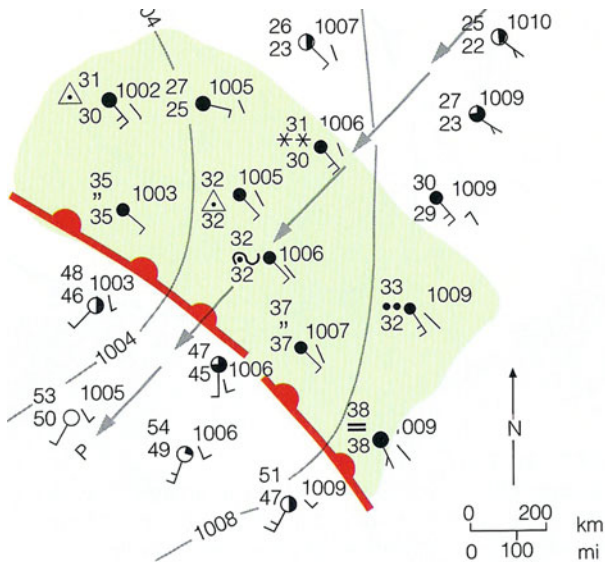


Fig. 2.10 Surface weather associated with a warm front

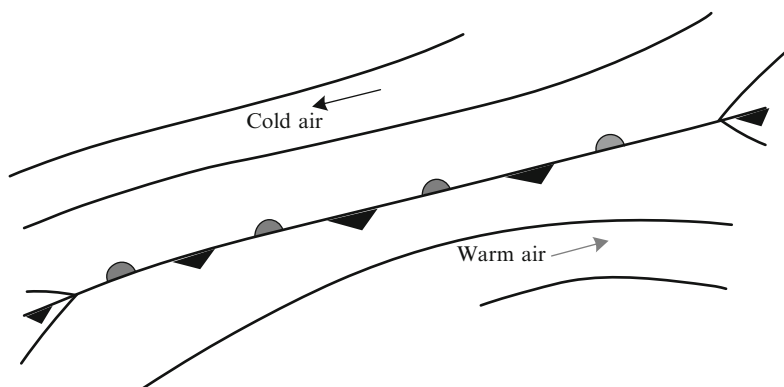
In a warm front the presence of warm air at the upper atmospheric layers prevents the mixing of air masses vertically. The presence of warm air masses that are rich in humidity and that move above cold air masses has as a result a volume increase in air mass and the formation of clouds and rain with gradual cooling of the air. At the warm fronts there is the formation of extended cloud layers and storms with the scavenging of chemical compounds which are inside them. Transport from the lower layer is not easy due to the absence of clouds.

The gradient of the warm front in relation to the surface is close to  $1/100$ . The warm air, since it is lighter, is moving faster than the cold air and is ascending above it along the warm frontal surface. During its ascent the warm air is getting colder adiabatically and after the saturation point starts the condensation process with the formation of extended cloud layers. The width of these cloud layers extends between 100 and 300 miles and its length may extend 1,000 miles. Since the gradient of the warm front is small, then the upward movement of the warm air masses occurs slowly. The warm air masses are characterized by stable conditions due to the uniformity of the temperature and relative humidity characteristics inside them, and the clouds which are formed are stratocumulus. These clouds are very extensive geographically and give rain over a range up to 2,000 miles which may be light or medium in intensity but has long duration. In special cases the warm air can be unstable so there is a possibility of formation of clouds with vertical development (storms).

The limit on the Earth's surface between the two air masses is depicted in the meteorological maps as a red line with semi-circles which show the movement direction of the warm front. In Table 2.4 are presented the main characteristics of

**Table 2.4** Characteristics of warm and cold fronts

Characteristics	Cold front	Warm front
Distribution of air masses	The cold air mass press the warm air mass	The warm air mass ascends above the cold air mass and pushes it away
Gradient of the frontal surface (in relation to the surface)	Large gradient ( $1/50$ )	Small gradient ( $1/100$ )
Weather phenomena	Intense weather phenomena (storms, strong winds, strong rain etc.)	Moderate weather phenomena (weak to moderate rain, moderate winds)
Band of weather phenomena	Close to 150 km (100 km front of the cold front and 50 km behind it)	400 km (or more) The weather phenomena occur before the warm front
Duration of the weather phenomena	Small duration	Large duration
Symbol	Solid blue line with triangles along the front showing its direction of movement	Solid red line with half circles along the front showing its direction of movement



**Fig. 2.11** Stationary front at a weather surface map

the warm and cold fronts. These characteristics can be modified to a larger extent since they are dependent on many parameters, such as the season, the thermodynamic characteristics of the air masses, the humidity content of the warm air mass and others.

#### 2.3.3.4 Stationary Fronts

When warm and cold air masses are in contact but are not moving and therefore neither of them is tending to displace the other, then the intersection of their dividing surface with the Earth's surface is called a stationary front (see Fig. 2.11).

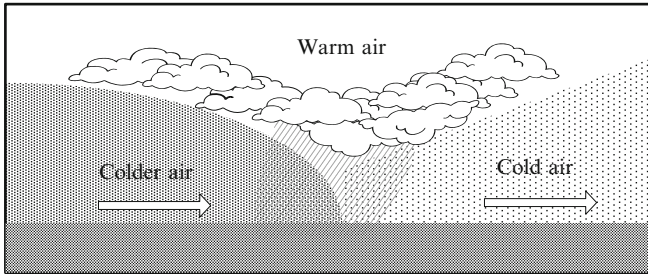
The wind is moving parallel to the stationary front and the weather conditions at these fronts are similar to the warm fronts. However, they have weaker intensity and can cover a larger area. The stationary fronts can remain for several days or weeks and their exact weather conditions are difficult to be forecast. In surface maps they are depicted with a solid line which has red arrows and blue semi-circles.

#### 2.3.3.5 Occluded Fronts

Cold fronts move with higher velocity in relation to warm fronts and as a result the warm region (the area between the cold and warm fronts) is reduced and finally the cold front catches up and overtakes a warm front with the formation of an occluded front. This can occur at the final stages of a wave cyclone. In the formation of an occluded front, three air masses are involved and their position vertically is dependent on their temperature difference (Fig. 2.12)

The occluded front is depicted with a line on which triangles (blue) and hemi-circles (red) denote the direction of the front movement.

The clouds that accompany an occluded front are dependent on the clouds which were present at the warm and cold fronts. Severe weather conditions may occur at



**Fig. 2.12** Vertical structure of an occluded front

the first stages of the formation of an occluded front due to the unstable air mass which is forced to move upward. However, this stage lasts only for a short time interval.

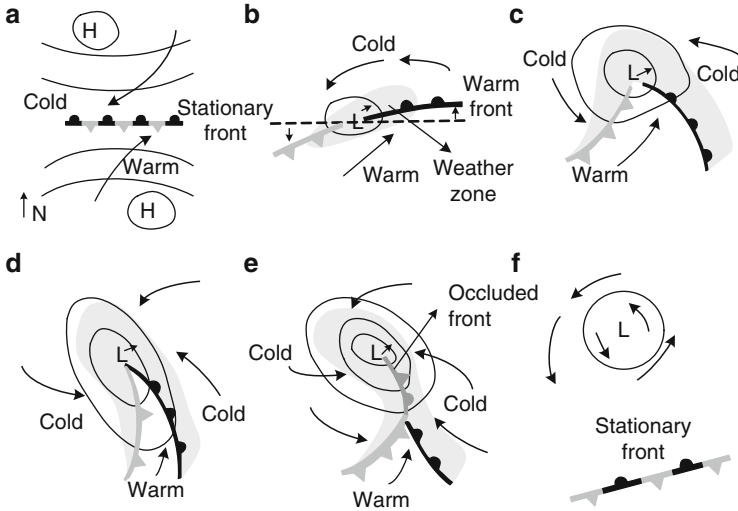
### 2.3.4 Wave Cyclone

The low barometric conditions at the middle latitudes are developed at regions where there is a formation of fronts. The fronts are formed due to the general atmospheric circulation close to the Earth's surface.

The process starts with a stationary front, where the cold air is located north and the warm south. A front has the properties of a wave. Due to the front tilt there is a formation of an initial oscillation at the intersection between the front and the surface. Thus the warm air forms a cavity inside the cold air. The pressure starts to drop at the top of the cavity and starts the formation of a cyclonic movement of air. The low pressure system which is formed and is accompanied with frontal movements (warm and cold front) is called a wave cyclone. When the wave cyclone is formed it is moving in an easterly direction, passing successively through stages as depicted in Fig. 2.13 until its dissolution (life cycle of a wave cyclone).

The first stationary front is divided into a warm and a cold front and the low system which accompanies it starts to deepen. The sector between the warm and cold fronts is called a warm sector. The whole system moves eastward and the pressure at the centre of the low system continues to decrease. The transport velocity of a wave cyclone is equal to the velocity of the geostrophic wind in the warm region.

A continuing pressure drop causes convergence inside the low system and consequently an upward flow. This upward flow forms due to adiabatic cooling of water condensation and other weather phenomena. The weather conditions are discussed in the sections related to warm and cold fronts. As the air of the warm sector ascends in conjunction with the quick arrival of the cold front, a narrow area of the warm sector develops. Then the cold front occupies the warm section and the two fronts are combined. The combination starts at the Earth's surface. The front which is formed from the above process is the occluded front. Finally, the barometric low gradually disappears.



**Fig. 2.13** A life cycle of a wave cyclone

## 2.4 Turbulence – Equations for the Mean Values

Turbulence characterizes the atmospheric boundary layer; but, due to the complex structure and variability of the layer, a deterministic description of its turbulence is difficult. A description of turbulence can, however, be constructed through its statistical properties. A suitable methodology is to divide the flux into turbulent and non-turbulent terms. An example is the calculation of the changes in small spatial and temporal changes, where the equations are expanded to average and instant values. This methodology is known in the literature as Reynolds analysis since it was developed by Osborne Reynolds.

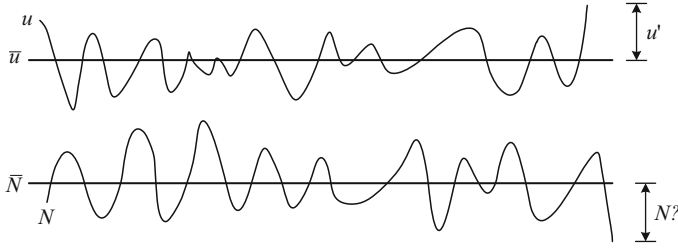
As an example, the gaseous number concentration can be divided as the sum of the average and instant values:

$$N = \bar{N} + N' \quad (2.10)$$

where,  $\bar{N}$  is the average gaseous concentration (molecules/(m<sup>3</sup> s)) and  $N'$  is the instant value. The average concentration is calculated with integration of time and volume:

$$\bar{N} = \frac{1}{\Delta t \Delta x \Delta y \Delta z} \int_t^{t+\Delta t} \left\{ \int_x^{x+\Delta x} \left[ \int_y^{y+\Delta y} \left( \int_z^{z+\Delta z} N dz \right) dy \right] dx \right\} dt. \quad (2.11)$$

The atmospheric flux is turbulent. Turbulent flux does not have specific forms but exhibits a random behavior in relation to time. These random changes of velocity give changes to the rates of change of momentum, heat and mass which are higher



**Fig. 2.14** Actual, average and instant values for the velocity and gaseous concentration. Every point in the horizontal axis depicts variations at specific time and space values

by several orders of magnitude in relation to the molecular diffusion. During turbulent flux there is a continuous conversion of the kinetic energy to internal energy. The energy source in turbulent dispersion is the shear flux in the flux field.

If  $u_i(t)$  is the instant value of the velocity of a molecule, then the average velocity value versus time is given by the expression

$$\bar{u}_i = \lim_{\tau \rightarrow \infty} \frac{1}{\tau} \int_{t_0}^{t_0 + \tau} u_i(t) dt. \quad (2.12)$$

Since the velocity is not exactly constant versus time, the average velocity can be expressed as

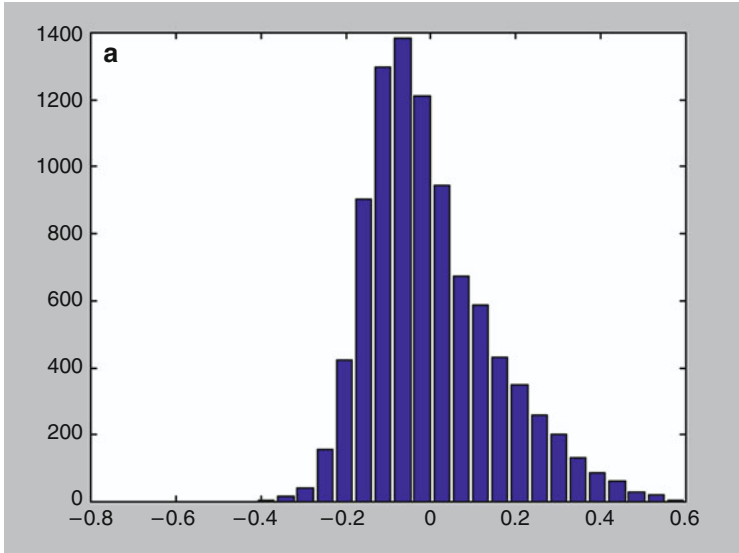
$$\bar{u}_i(t) = \frac{1}{\tau} \int_{t-\tau/2}^{t+\tau/2} u_i(t') dt'. \quad (2.13)$$

Figure 2.14 shows the time series of velocity  $u$  and the gaseous concentration  $N$ , depicting the difference between average and instant values.

## 2.5 Statistical Properties of Turbulence

Turbulence is a characteristic of the atmospheric boundary layer and due to its non-deterministic nature its description is formulated with the help of statistics. In this section statistical methods for the study of turbulence are examined.

Turbulence has a spectrum analogous to the colour spectrum of light after passing through prism. Similar analysis is performed for the description of the colour spectrum and turbulence such as the contribution of different components of flux to the total turbulent kinetic energy. Figure 2.15 shows an example of the function of the probability density for temperature in the atmospheric boundary layer. It is important to understand that variations of different components in the atmosphere make necessary their statistical evaluation.



**Fig. 2.15** Typical form of the function of the probability density for temperature fluctuations in the atmospheric boundary layer (x-y plane at  $z = 389$  m). The y-axis denotes frequency of occurrence, and the units of the x-axis are degrees Kelvin. Mean temperature is 301 K (Adapted from Housiadas et al., 2004)

The function for the mean value constitutes one of the main functions which will be examined. There are several ways to examine the mean value of a function. These include mean values of time  ${}^t(\bar{\phantom{x}})$ , space  ${}^s(\bar{\phantom{x}})$ , and from the ensemble  ${}^e(\bar{\phantom{x}})$  (Stull 1997). Therefore for a variable  $A(t,s)$  that is function of time ( $t$ ) and space ( $s$ ), the following expressions for discontinuous and continuous conditions can be used:

$${}^t\bar{A}(s) = \frac{1}{N} \sum_{i=0}^{N-1} A(i, s) \quad \text{or} \quad {}^t\bar{A}(s) = \frac{1}{P} \int_{t=0}^P A(t, s) dt, \quad (2.14)$$

$${}^s\bar{A}(t) = \frac{1}{N} \sum_{j=0}^{N-1} A(t, j) \quad \text{or} \quad {}^s\bar{A}(t) = \frac{1}{S} \int_{s=0}^S A(t, s) ds, \quad (2.15)$$

$${}^e\bar{A}(t, s) = \frac{1}{N} \sum_{i=0}^{N-1} A_i(t, s) \quad \text{or} \quad {}^e\bar{A}(t) = \frac{1}{E} \int_{e=0}^E A(t, e) de. \quad (2.16)$$

A statistical criterion for the variation of data around its mean value is the dispersion coefficient which is defined as

$$\sigma_A^2 = \frac{1}{N} \sum_{i=0}^{N-1} (A_i - \bar{A})^2. \quad (2.17)$$

The dispersion coefficient is a good measure of the variations inside the boundary layer. Another coefficient which is used more often for groups of measurements can be defined as

$$\sigma_A^2 = \frac{1}{N-1} \sum_{i=0}^{N-1} (A_i - \bar{A})^2, \quad (2.18)$$

where the instantaneous variations can be written as

$$a'_i = A_i - \bar{A}. \quad (2.19)$$

Finally we can write as a result,

$$\sigma_A^2 = \frac{1}{N} \sum_{i=0}^{N-1} a_i'^2 = \overline{a'^2}. \quad (2.20)$$

In addition, a coefficient can be defined for the intensity of turbulence for an average velocity value  $\bar{U}$  as

$$I = \sigma_U / \bar{U}. \quad (2.21)$$

The covariance coefficient denotes the correlation between variable A and B:

$$\text{covar}(A, B) \equiv \frac{1}{N} \sum_{i=0}^{N-1} (A_i - \bar{A}) (B_i - \bar{B}). \quad (2.22)$$

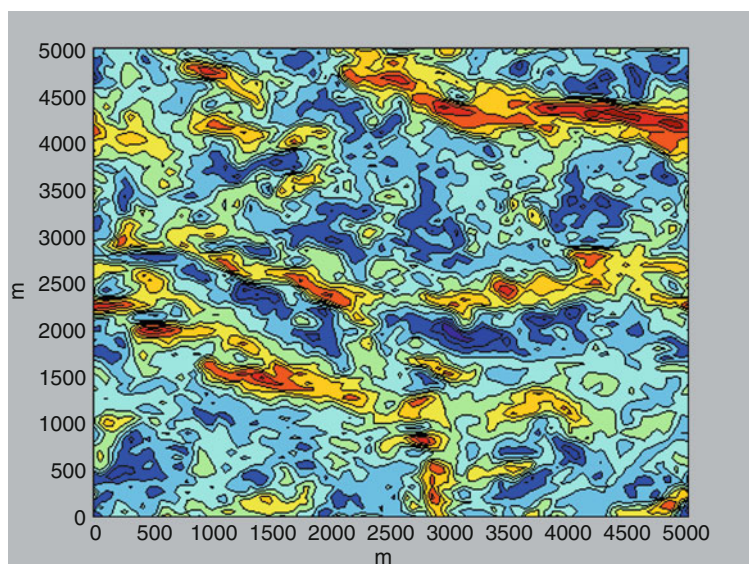
Using the Reynolds methodology ( $\overline{(AB)} = \bar{A}\bar{B} + \overline{a'b'}$ ) for the average value, then the following expression is derived for the covariance (covar(A,B)):

$$\text{covar}(A, B) \equiv \frac{1}{N} \sum_{i=0}^{N-1} a'_i b'_i. \quad (2.23)$$

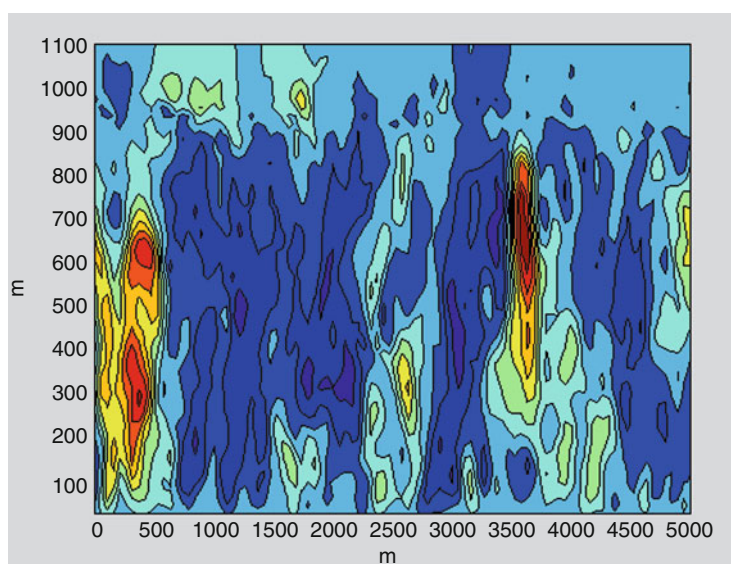
As an example for the covariance coefficient, we examine the case that the coefficient A is the temperature (T) and B is the vertical velocity component w. On a warm summer day the warmer air ascends (positive T' and w') and the colder air descends (negative T' and w'). This denotes that the product w' × T' is positive and therefore the temperature and vertical velocity components change co-instantaneously. Figure 2.16 shows the vertical fluctuations of the velocity at the surface x – y for a typical situation of the atmospheric boundary layer, whereas, Fig. 2.17 shows the vertical fluctuations of the velocity at the surface y – z.

The normalized covariance ( $r_{AB}$ ) can be defined as

$$r_{AB} \equiv \frac{\overline{a'b'}}{\sigma_A \sigma_B}. \quad (2.24)$$



**Fig. 2.16** Instantaneous  $w$ -fluctuations (m) in the  $x$ - $y$  plane at  $z = 190$  m (middle of domain) (Adapted from Housiadas et al., 2004)



**Fig. 2.17** Instantaneous  $w$ -fluctuations (m) in the  $y$ - $z$  plane at  $x = 2,500$  m (middle of domain). (Adapted from Housiadas et al., 2004)

The coefficient  $r_{AB}$  has values which range between +1 and -1. When two variables change in the same manner, then  $r_{AB} = 1$ . When the variables change in opposite ways, then  $r_{AB} = -1$ .

The kinetic energy of mass  $m$  is given by the expression  $E_k = 1/2 m v^2$ , where  $v$  is the velocity. The kinetic energy per unit mass is written as  $E_k/m = 1/2 v^2$ . The kinetic energy of flux can be divided into two terms, one which is related to the mean velocity and one which is related to the turbulence. Therefore the expression for the mean kinetic energy can be written as

$$\frac{MKE}{m} = \frac{1}{2} (\overline{U^2} + \overline{V^2} + \overline{W^2}), \quad (2.25)$$

where  $\overline{U}$ ,  $\overline{V}$ ,  $\overline{W}$  are the three components of the mean velocity.

The expression for the turbulent kinetic energy can be written as

$$\frac{TKE}{m} = \frac{1}{2} (\overline{u'^2} + \overline{v'^2} + \overline{w'^2}), \quad (2.26)$$

where,  $u'$ ,  $v'$  and  $w'$  are the three components of the instantaneous velocity.

Furthermore we study the flux concept in the atmosphere, which is defined as the transport of a quantity per unit surface and unit time. In the boundary layer there is usually a study of flux of mass, heat, humidity, momentum and pollutants. As an example for the mass we define the pollutant flux ( $\tilde{M}$ ) which is expressed in units  $\left[\frac{Kg_{air}}{m^2 s}\right]$ . The kinematic flux is defined as  $M = \frac{\tilde{M}}{\rho_{air}}$  (units  $\left[\frac{m}{s}\right]$ ). Furthermore the vertical kinematic heat flux due to turbulent flux (vertical kinematic eddy heat flux) is defined as  $w'\theta'$  where  $\theta$  is the temperature. Another quantity that is important in the study of atmospheric flux is stress, which is actually a force that can produce a deformation of a body. Pressure is a kind of stress which is applied in fluids under equilibrium. The Reynolds stress for a fluid in turbulent movement is given by the expression

$$\tau_{Reynolds} = -\rho \overline{u'w'} \quad (\text{Reynolds stress}) \quad (2.27)$$

or can be expressed as

$$\tau_{ij} = \mu \left( \frac{\partial U_i}{\partial x_j} + \frac{\partial U_j}{\partial x_i} \right) + \left( \mu_B - \frac{2}{3} \mu \right) \frac{\partial U_k}{\partial x_k} \delta_{ij}, \quad (2.28)$$

where,  $\mu_B$  is the viscosity coefficient and  $\mu$  is the dynamic viscosity coefficient.

The tensor of the Reynolds stress is symmetric and is given as

$$\begin{bmatrix} \overline{u'u'} & \overline{u'v'} & \overline{u'w'} \\ \overline{u'v'} & \overline{v'v'} & \overline{v'w'} \\ \overline{u'w'} & \overline{v'w'} & \overline{w'w'} \end{bmatrix} \quad (2.29)$$

where  $u$ ,  $v$ ,  $w$  are the air velocities in the three directions.

A typical value of the kinematic coefficient of the Reynolds stress in the atmosphere is  $0.05 \text{ m}^2/\text{s}^2$ . The Reynolds stress is a characteristic property of the flux and not of the medium.

In addition the Reynolds stress at directions  $x$ ,  $y$  and  $z$  can be expressed as

$$\tau_{xz} = -\bar{\rho} \overline{u'w'_s}, \quad (\text{viscous stress}) \quad (2.30)$$

$$\tau_{yz} = -\bar{\rho} \overline{v'w'_s}, \quad (2.31)$$

with a total Reynolds stress

$$|\tau_{\text{Reynolds}}| = [\tau_{xz}^2 + \tau_{yz}^2]^{1/2}. \quad (2.32)$$

Finally the friction velocity  $u_*$  can be written as

$$u_*^2 \equiv [\overline{u'w'_s}^2 + \overline{v'w'_s}^2]^{1/2} = |\tau_{\text{Reynolds}}|/\bar{\rho}. \quad (2.33)$$

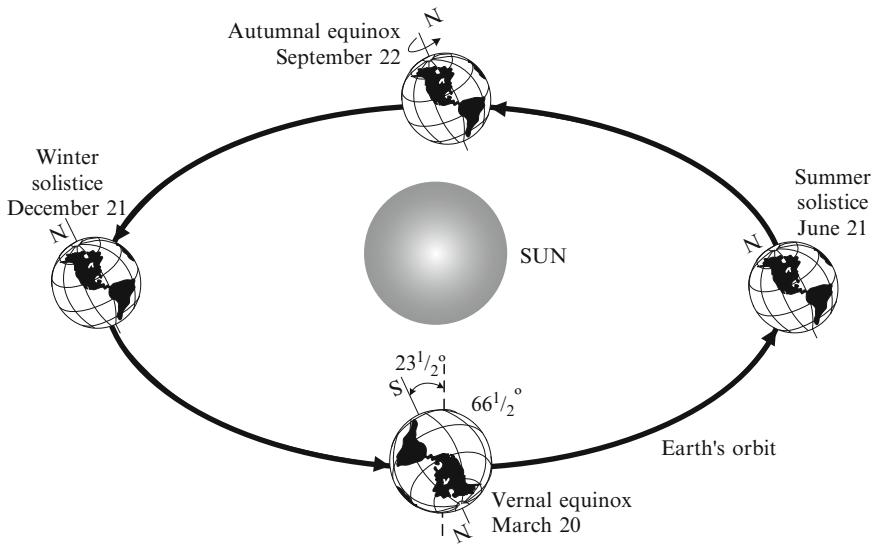
## 2.6 Atmospheric Temperature

### 2.6.1 Temperature Season Variability

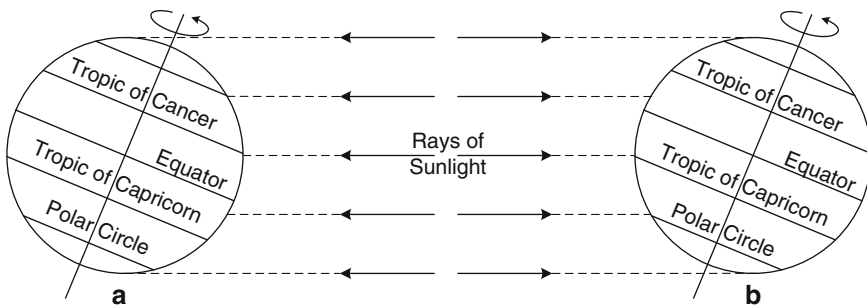
The Sun is the largest heat source in our solar system and of course also for the Earth. Heat is an energy form that is dependent on the composition of materials and can be moved to other bodies or can be transferred to other energy forms.

The Earth spins on its own axis (with a period of one day – 24 h) and at the same time revolves with a velocity of thousands of kilometers per hour around the sun with elliptical trajectory (with a period close to 365 days) (Fig. 2.18). The direction of rotation is anticlockwise and has a velocity of several 100 kilometers per hour. This is the reason that sun, moon and stars rise in the east and decline in the west.

Since the Earth performs an elliptical orbit around the sun, the distance between Sun – Earth changes during the year. Earth is closer to the Sun in January (distance of 147 million km) than in June (distance of 152 million km). At the Northern Hemisphere the average temperature in June is higher than that of January and thus the change of seasons is determined by the quantity of the Sun's energy which reaches the earth. The quantity of radiation from the Sun which reaches the earth's surface is determined from the angle that radiation strikes the surface and the time period that the sun radiates at a specific latitude.



**Fig. 2.18** Earth's orbit around the Sun. During this movement the Earth is rotated around an axis which has an angle of  $23\frac{1}{2}^\circ$  from its vertical axis. The rotation axis is constant during the year. The result is that in June, when the northern hemisphere is turned towards the Sun, there is more direct light and longer duration of the day there. This results in warmer weather, compared to December when the northern hemisphere is turned away from the Sun.



**Fig. 2.19** Earth's direction (a) 21st June (b) 21st December

Figure 2.19 shows also the earth's direction in relation to the Sun's radiation during 21st June and 21st December. On 21st June the Sun's radiation strikes the earth vertically at  $23\frac{1}{2}^\circ$  north (N) (Tropic of Cancer). On 21st December the darkness lasts for 24 h for all areas above  $66\frac{1}{2}^\circ$  north (N) and  $66\frac{1}{2}^\circ$  south (S).

When the Sun's radiation penetrates the atmosphere, a part of it is absorbed or scattered from the atmospheric gasses, whereas another part is reflected from clouds. This means that the greater is the thickness of the atmosphere which the Sun's radiation has to penetrate, the higher is the possibility that the radiation will be absorbed or scattered. This occurs during summer and high latitudes such as

in the Scandinavian countries. The Sun in these countries is never very high on the horizon and therefore the radiation must penetrate a quite thick atmospheric layer before it reaches the Earth's surface. Also due to high cloud content of the atmosphere the radiation is scattered effectively before it reaches the ground.

The seasons are determined from the quantity of the Sun's radiation which reaches the planet, which is determined from the duration of the day and the incident angle of the Sun's radiation. As a result the higher latitudes lose more radiation (from the emitted radiation at the higher wave length of the Earth's surface) than what they receive (short wave length of the Sun's radiation). On the contrary, at lower latitudes there is a higher quantity of Sun's radiation which reaches the Earth than is emitted from it. Due to the global wind circulation that results from this temperature difference there is a transport of heat from the lower to higher latitudes.

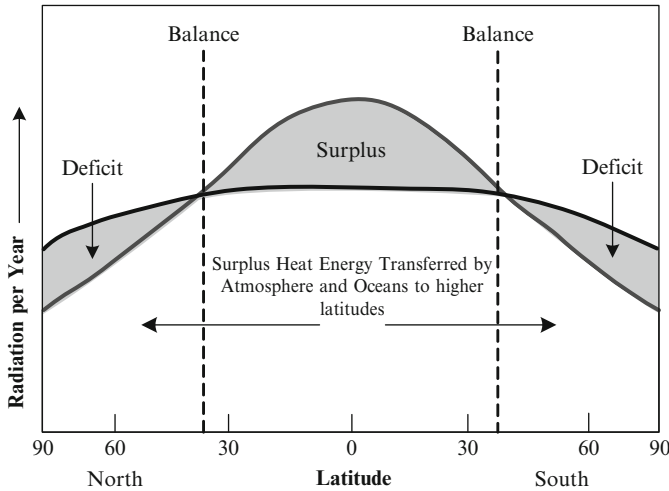
As shown in Figs. 2.18–2.19, on 21st June the Sun's radiation has a maximum at the Earth's surface at latitude  $30^{\circ}\text{N}$ . This day the Sun is located above the latitude  $23\frac{1}{2}^{\circ}\text{N}$  (Tropic of Capricorn). The reason that the maximum of incident energy is at latitude  $30^{\circ}\text{N}$  and not at latitude  $23\frac{1}{2}^{\circ}\text{N}$  is due to two reasons. First, the duration of day at the area of  $30^{\circ}\text{N}$  is larger than the area  $23\frac{1}{2}^{\circ}\text{N}$  on 21st June. Secondly the area close to the point  $30^{\circ}\text{N}$  is characterized by desert areas, clear sky and dry air, whereas at the area  $23\frac{1}{2}^{\circ}\text{N}$  the climate is more humid with clouds which reflect the Sun's radiation.

After the 21st of June the sun is lower at noon in the sky and the summer days become shorter. With the beginning of September there is the start of autumn. On September 22nd the Sun is located above the equator. On this date, except at the poles, the day and night have the same duration (autumnal equinox). At the north pole the Sun appears on the horizon for 24 h and then disappears from the horizon for 6 months. At other latitudes in the north hemisphere after the 22nd September the Sun at noon appears gradually lower in the sky and the day lasts fewer hours.

On 21st December, which is 3 months after the autumnal equinox, the north hemisphere is directed farther from the Sun compared to the previous period. The nights are long and the days are short. This is the shortest day of the year (winter solstice) and it is the astronomical start of winter. This day the Sun shines directly above latitude  $23\frac{1}{2}^{\circ}\text{S}$  (Tropic of Capricorn). It is located at its lower position at the middle of the day and its radiation passes through a large portion of the atmosphere and affects a large portion of the Earth's surface.

On 20th March there is the astronomical start of spring and it is the vernal equinox. This day the Sun shines directly above the equator, whereas at the north pole the Sun appears on the horizon after being 6 months absent. The following period the days last longer and there is warmer weather in the north hemisphere.

After 3 months, 21st June, the sun's radiation reaches a maximum. It is obvious that even though the sun's radiation is more intensive during June, the warmest period appears a few weeks later during July or August. The reason is that during June the outgoing radiation from earth is smaller than the incoming and there is no thermal equilibrium. When a thermal equilibrium is achieved, there is the highest temperature in the atmosphere and this is achieved a few weeks after the 21st June. The same situation occurs during winter when the outgoing energy from the



**Fig. 2.20** The average annual incoming solar radiation (grey line) absorbed by the earth and the atmosphere compared with the average emitted infrared radiation from the earth and the atmosphere (Adapted from Ahrens 1994)

atmosphere is higher than the incoming and the temperature decreases. Since the outgoing radiation is higher than the incoming then for a few weeks after 21st December there is the coldest period mainly during January and February.

Whereas in the whole planet there is energy equilibrium between the absorption and emission of solar radiation, this is not valid at every latitude. Figure 2.20 shows the energy equilibrium at different latitudes.

The temperature is expressed in degrees, which are subdivisions of thermal scales. The best known and most used are the Celsius centigrade scale ( $^{\circ}\text{C}$ ) and the Fahrenheit scale ( $^{\circ}\text{F}$ ) which is used mainly in the United States and finally the Kelvin scale (K) which is used mainly for scientific purposes. Thermal energy is observed in the system Earth – atmosphere with the following forms:

- As absolute heat which can be measured with the help of thermometers and
- As latent heat which occurs during specific physical processes related to the phase changes of water (evaporation, condensation etc.).

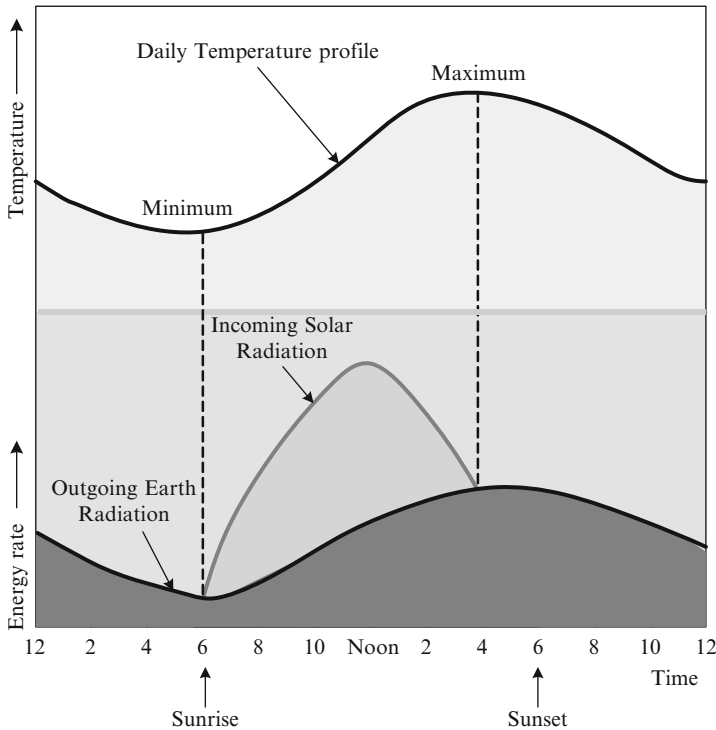
As discussed earlier, the main source of heat for the Earth and its atmosphere is the Sun, since the energy from the Earth's interior is negligible. It has been calculated that if there was no contribution from the Earth's interior its mean temperature would be reduced by less than  $0.1^{\circ}\text{C}$ . Therefore the short length Sun's radiation affects the planet's temperature and controls, together with the long length radiation, the temperature on the Earth's surface. The processes of heat exchange (between warm and cold regions) at the lower atmospheric layers due to uneven heating of the earth's surface are the main reasons for the formation of weather phenomena inside the atmosphere.

### 2.6.2 Temperature Daily Variability

The intensity of the Sun’s radiation which is received from the Earth’s surface is dependent on the Sun’s position. During a sunny day the radiation intensity has a simple variation with maximum close to noon. During a day without clouds and turbulent atmosphere the air temperature varies with the minimum a few minutes after the Sun’s rise and maximum 2–3 h after the noon hour. As shown in Fig. 2.21 during a normal day without clouds the upward part of the daily variation of the temperature has higher gradient compared to the downward part, since during the night there is no great loss of thermal radiation.

During the daily temperature variation in air the upper layer of the Earth’s surface absorbs and emits the Sun’s radiation, which is responsible for the temperature changes. During the night there is often a temperature inversion with an increase of the temperature versus height since there is cooling of the Earth’s surface with the emission of electromagnetic radiation with large wave length.

The difference between the maximum and minimum temperature during a day is called daily thermal width. This is larger above the mainland and smaller above the



**Fig. 2.21** Daily changes of the temperature and incoming and outgoing radiation. When the incoming Sun’s energy is higher than the outgoing energy then the air temperature increases. Contrary when the outgoing energy is higher than the incoming then the air temperature decreases. (Adapted from Ahrens 1994)

seas due to the larger heat capacity of water. Its value decreases gradually from equator to poles as the latitude increases.

In the northern hemisphere, over a year's time, the temperature of air in temperate regions shows a simple variation. The maximum occurs most of the time above mainland during July and above sea during August. The minimum is observed above mainland during February and above sea during March.

The difference between the mean temperature of the warmest and coldest month of a year is called annual thermal width. This is larger above mainland and smaller above sea and increases from the equator to the poles.

The first classification of climate is based on the annual thermal width:

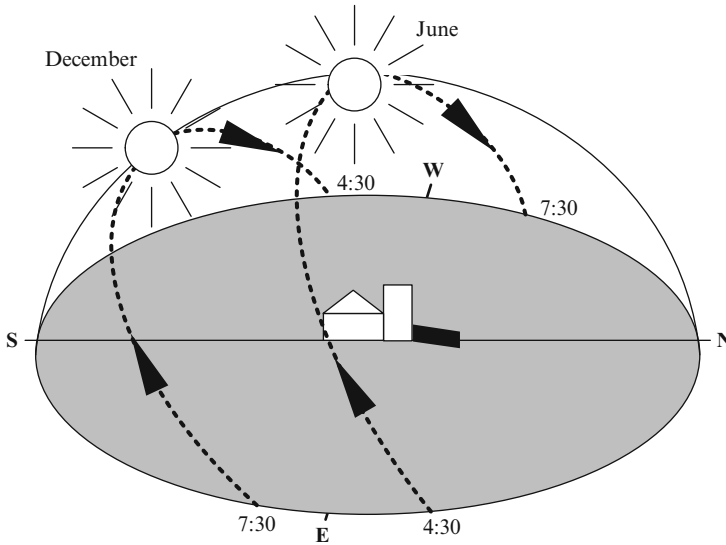
- Tropical moist, when the annual width is smaller than 10°C.
- Temperate, when the annual width is between 10°C and 20°C and
- Moist mid-latitude climates with severe winters, when it is larger than 20°C.

A detailed characterization of climate is demonstrated by the Köppen classification system (Ahrens 1994).

The air temperature is the most important climatic component and the most important parameter for climate classification. Meteorologists and climatologists examine temperature values at different elevations inside the atmosphere. When someone refers to air temperature, they mean the temperature in shadow inside a special shelter (meteorological cage) and at height 1.5–2.0 m above ground.

For mainly climatological reasons the air temperature in a given location can be described with the following parameters:

- Average daily temperature ( $\bar{T}_{\text{day}}$ ) which is defined from the expression  $\bar{T}_d = \frac{1}{24} \sum_{i=1}^{24} T_{h(i)}$  where  $T_{h(i)}$  is the hourly value ( $i = 1, 2, 3, \dots, 24$ ). This expression is used when the meteorological station has the capability of hourly temperature measurements. The expression which is used by the World Meteorological Organization for the calculation of  $T_d$  is:  $\bar{T}_d = \frac{1}{4}(T_{06} + T_{12} + 2T_{18})$ , (where 06, 12 and 18 are the time in UTC – Universal Time Coordination). The use of this expression is adopted in cases where there is need to have direct comparison between the meteorological values in an extensive geographical area. For climatological use the data have to be extended for at least a period of 30 years.
- Absolute maximum ( $T_{\text{max}}$ ) and minimum ( $T_{\text{min}}$ ) value of the air temperature which occurs during a whole day (24 h).
- Average monthly temperature ( $\bar{T}_{\text{mo}}$ ) which can be calculated from the expression  $\bar{T}_{\text{mo}} = \frac{1}{v} \sum_{i=1}^v \bar{T}_{d(i)}$ , where  $v$  is the number of days during the month.
- Daily thermal width. This is defined as the difference between the maximum and minimum temperature value (M.T.V.) during 24 h:  $\text{MTV} = T_{\text{max}} - T_{\text{min}}$ .
- Yearly temperature width (Y.T.W.). This is defined as the difference between the average air temperature of the coldest month and the average temperature of the warmest month during the year:  $\text{Y.T.W.} = \bar{T}_{\text{mo(warm)}} - \bar{T}_{\text{mo(cold)}}$



**Fig. 2.22** Variability of the Sun's orbit at mean geographical widths at the northern hemisphere (Adapted from Ahrens 1994)

Figure 2.22 shows the trajectory of the Sun at middle latitudes in the northern hemisphere during a year. During winter the sunrise starts at the south-east and the sunset occurs at south-west. During summer the sunrise occurs in the northeast and the sunset in the northwest. Consequently a house with windows facing south receives more light than a house with windows facing north. The same applies more intensively for mountainous regions. Regions which are oriented to the south receive more light and as a result tend to be warmer and drier than regions which are oriented to the north. This has a direct effect on the flora of the region. Vineyards which are located at areas oriented to the south produce a better quality wine. On the contrary plants which withstand cold are located at areas oriented to the north. The architecture of houses is also affected from the position of the sun and considerable benefits for energy consumption are derived from an appropriate orientation of houses and the rational use of windows (bioclimatic architecture). Another example is the operation of ski resorts which have a north orientation.

### 2.6.3 Heating of the Earth's Surface and Heat Conduction

Radiation from the sun is the main heat source for the Earth's surface. The heating of the surfaces and their temperature is dependent on a number of parameters such as:

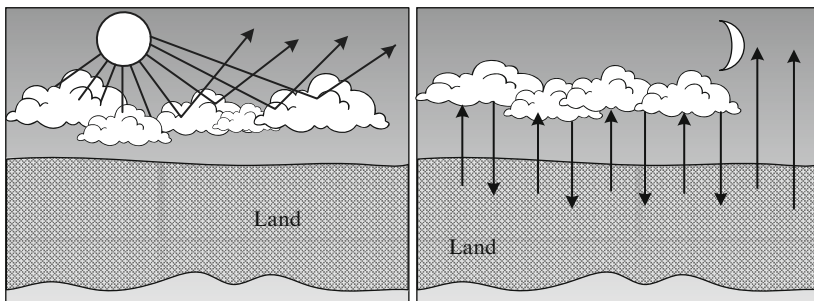
- The special heating (**heat capacity**) of a surface. By the term heat capacity is meant the quantity of heat that is needed to increase the temperature of one gram

of a given material by  $1^{\circ}\text{C}$ . Since the heat capacity of water is twice that of the ground, then more heat is needed in order to increase the temperature of a water surface by  $1^{\circ}\text{C}$  compared to the ground. As a result the ground becomes heated (or cooled) more quickly compared to the sea. In comparison with the sea, the ground is warmer during the day but it is cooler during the night.

- The **absorbency** of a surface. Everybody that receives a quantity of radiation absorbs part of its energy. The percentage of absorption is dependent on the body's nature and the radiation.
- The **reflectivity** of a surface. If the total amount of the sun's radiation is reflected from a surface, there is no absorption or transformation to thermal energy. The surfaces of snow and water have high reflectivity and are not heated, such as surfaces with low reflectivity (e.g. cultivated areas, dense jungle etc.).
- The **conductivity** of a surface. The ocean streams transport large quantities of heat (inside oceans) through the water movement. With this process the sea is heated to a larger depth compared to the mainland.
- The **cloud cover** of a surface. A factor which plays a significant role in the temperature of surfaces is the cloud cover, which during the day does not allow penetration of the Sun's radiation to the Earth's surface (Fig. 2.23). This results in a reduction of the Earth's temperature. Therefore the air which is in contact with the Earth's surface will receive lower heating during the day. During the night the cloud cover results in the opposite phenomenon since it does not allow part of the thermal energy which it receives from the Earth to escape to space. The atmosphere under clouds experiences lower cooling and therefore higher temperatures occur.

The thermal energy is spread from one body to the next or re-distributed inside a body in different ways. Some of these mechanisms are:

- **The radiation.** All bodies emit energy in the form of electromagnetic radiation. Higher temperatures result in smaller wave length of the radiation. Consequently the wave length of the Sun's radiation is smaller than the re-emitted radiation from the Earth's surface which is much colder than the Sun's surface.



**Fig. 2.23** The cloud cover reduces the warming of the Earth's surface during day and its freezing during the night

- **The conductivity.** The thermal energy can be conducted inside a body or from one body to the other with contact through conductivity. For example iron is a good conductor of heat, whereas the wood or air are poor heat conductors. An air molecule which is in contact with the Earth's surface is heated through conductivity. This is an important factor for creation of the weather phenomena.
- **The heat transfer.** A moving body of air transports also its thermal energy. The heat transfer occurs vertically and horizontally inside the atmosphere and this process is the most important for the weather phenomena.
  - a. **Vertical heat transfer:** An air mass which is heated on the Earth's surface starts to expand and then becomes less dense and ascends. With its uplift it expands adiabatically and transfers its thermal energy at upper atmospheric layers.
  - b. **Horizontal heat transfer:** An air mass moves horizontally to fill the gap of air which is created from the vertical movement of another air mass. This air mass which is moving horizontally transports together its thermal energy and humidity.

### ***2.6.4 Distribution of Temperature in the Air***

The temperature distribution of air above a region (small, large or even above the whole globe) is described with isothermal curves (lines with the same temperature along them).

The most important factors which control the temperature distribution in air are:

- The season and the latitude
- The distribution of land and sea
- The vegetation and general nature of the surface
- The elevation
- The slope of the Earth's surface
- The existence of snow or ice on the surface
- The cloud cover
- The prevailing winds and
- The sea streams.

Each of these parameters acts in a different manner and therefore the air temperature does not decrease smoothly from the equator to the poles. The highest temperatures are not observed at the equator but at latitudes  $10^{\circ}$ – $20^{\circ}$  south and north from it. This is due to the fact that at the equator there are extensive clouds and rainfall.

An important factor is the geographical distribution of land and sea. During summer the land is warmer than the sea. The temperature increases mainly above land. During winter the land is colder than the sea and therefore lower temperatures

are observed above land at higher latitudes (e.g. North Canada, Siberia, Greenland). Winds have a great influence on temperature distribution at different locations. Therefore the west winds in northern temperate zones transport warm ocean air masses to the west regions of Europe and America and cold air masses to the east regions of America and Asia.

The constant ocean currents finally influence considerably the temperature distribution. These are moving to the poles carrying warm water to colder regions, whereas the currents moving to the equator transport cold water masses to warmer regions.

## 2.7 Humidity in the Atmosphere

The term humidity refers to the water vapour which is contained in the atmosphere at a specific time. The air humidity is a very important parameter since it determines the cloud formation and the rain formation.

The atmosphere and especially troposphere contains water vapour at variable quantities which come mainly from water evaporation. The quantity of water vapour that the air contains is specific and is dependent directly on the air temperature.

When the air contains the maximum quantity it is called saturated. When only a part of the maximum quantity is contained, then it is called unsaturated. The term humid air is often used when there is an elevated amount of water vapour in the atmosphere. The term dry air is used in the absence of water vapour.

The water in the atmosphere is not only in the vapour phase but it exists also in the liquid phase (cloud, rain) and in the solid (snow, drizzle). It is known that in the atmosphere there is a water cycle. The water enters the air through evaporation from all the water surfaces and especially from oceans. Therefore huge quantities of water evaporate from the Earth's surface and are moved higher in the atmosphere and are condensed forming extensive cloud layers. Further, the clouds are transported to other regions and through wet deposition the water comes back to the Earth's surface.

Taking into account that  $\frac{2}{3}$  of the planet are covered by water, millions of tons of water are evaporated daily and are moved to the upper troposphere in the form of water vapour. The 84% of water vapour originates from oceans whereas the other 16% originates from lakes, rivers, wet surfaces, vegetation and the expiration of animals.

Every year the planet receives from water precipitation (e.g. rain, snow)  $400 \text{ km}^3$  of water. An equal quantity is introduced to the air through evaporation. Of the total water precipitated on the Earth's surface, close to  $100 \text{ km}^3$  precipitate on land and the other  $300 \text{ km}^3$  on oceans and water surfaces. Evaporation of  $400 \text{ km}^3$  water into the atmosphere during 1 year requires  $3 \times 10^{29}$  cal. This energy corresponds to 23% of the total energy which the Earth receives during a year from the sun.

The water in the atmosphere can change from one phase to another and for the water condensation the atmosphere has to be saturated (relative humidity equal to 100%). However in specific cases there is no condensation even at relative humidity

above 100%. This is due to the absence of a sufficient number of available condensation nuclei in the atmosphere. Without these nuclei, a relative humidity close to 420% is required in order for condensation to occur. Existence of sufficient condensation nuclei will lower the relative humidity required for condensation to 100%. With the presence of nuclei of sodium chloride the necessary relative humidity for condensation is close to 97%, whereas for sulphur oxide or phosphate oxide nuclei the relative humidity drops to 80%. Therefore in industrial regions the dense fog is a consequence of the existence of a large number of condensation nuclei. To the contrary, the number of condensation nuclei drops at higher elevations in the atmosphere.

As the water vapour condenses onto condensation nuclei, droplets or ice crystals form and with time their size increases. The formation of droplets or ice crystals is dependent on the temperature and pressure. At higher elevations where the temperature is far below 0°C and the pressure is also low, the water exists in the liquid phase. This is an unstable condition which is called super critical melting. When this unstable condition of water is disturbed by the passage of an aircraft, there is an immediate icing of the water droplets above the airframe and the wings which has consequences for the flight. Water at super critical melting exists at temperatures up to -15°C and, though less often, at -40°C. The most dangerous region is between 0°C and -15°C.

Generally the percentage of water vapour which comes from water evaporation from the soil is about 15% of the total water vapour which exists in air. The other 85% comes from the oceans. It is surprising to think that if the total amount of water in the atmosphere precipitated simultaneously onto the Earth's surface, it would cover the whole surface of the planet to a height of 2.5 cm.

Figure 2.24 shows the average spatial distribution of rainfall in Europe during the period 1940–1995. Lower values are observed in Southern Europe.

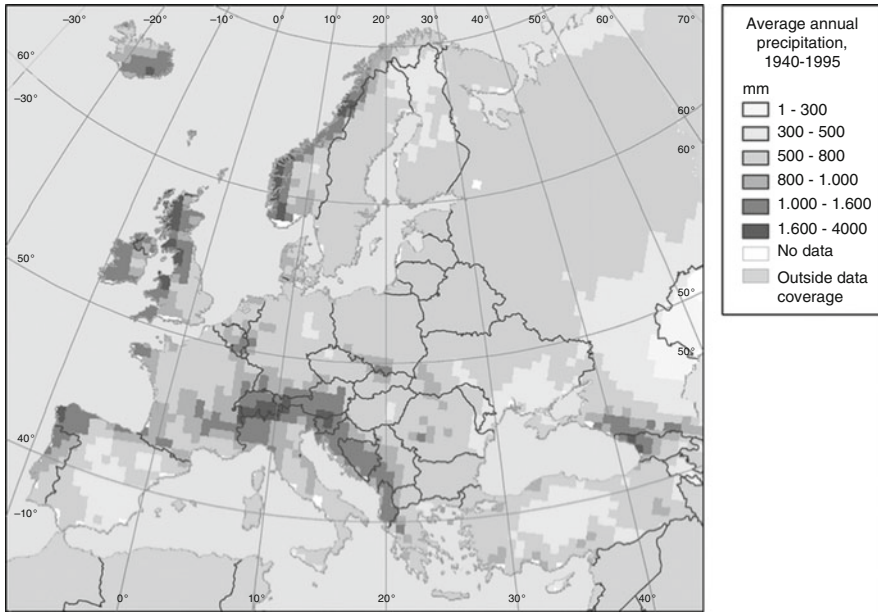
### 2.7.1 *Mathematical Expressions of Humidity in the Atmosphere*

The term humidity refers to the water content inside air. We present here some different methodologies that can describe this phenomenon.

#### 2.7.1.1 **Absolute Humidity (B)**

Absolute humidity  $\beta$  is defined to be the ratio of a mass of water vapour to the air volume in which the mass is contained. Absolute humidity denotes the density of water inside an air volume and usually is expressed as grams of vapour inside a cubic meter of air. For example if the water vapour inside a cubic meter of air weights 25 g, then the absolute humidity  $\beta$  of air is 25 g/m<sup>3</sup>.

The air volume changes with fluctuations of its elevation due to differences in air pressure. This results in changes in absolute humidity even though the quantity of



**Fig. 2.24** Map of the spatial distribution of precipitation (mm) in Europe in the period 1940–1995

water vapour inside the volume remains constant. For this reason the term absolute humidity is not often used in atmospheric sciences.

### 2.7.1.2 Specific Humidity ( $Q$ )

Specific humidity  $q$  is the ratio of the water vapour of the mixture divided by the total air mass and is expressed as water vapour grams per gram of humid air.

### 2.7.1.3 Mixing Ratio ( $R$ )

Mixing ratio is the ratio of water vapour to the mass of dry air.

### 2.7.1.4 Relative Humidity ( $RH$ )

Relative humidity of the atmospheric air is called the ratio of the water vapour mass which is contained in a specific air volume to the mass of the water vapour in the same volume under saturation conditions at the same pressure and temperature conditions. The relative humidity is actually the ratio of water vapour pressure to its equilibrium pressure at the same temperature. Therefore the relative humidity can be expressed as

$$RH = 100 \frac{p_{H_2O}}{p^o_{H_2O}}, \quad (2.34)$$

where the multiplication by 100 occurs since RH is expressed in percentage.

The relative humidity is an important parameter which is involved directly in the daily life of humans. An example is the study of the transfer of cold outdoor air to indoors. Dry arctic air incorporates small quantities of water vapour. Saturated air at temperature  $-25^\circ\text{C}$  includes only 0.5 g of water vapour per 1 kg air. When this air is introduced indoors at temperature  $20^\circ\text{C}$ , then its capacity to incorporate water vapour increases by 29 times to the value 14.7 g/Kg. This results in relative humidity indoors of

$$RH = 100 \times \frac{0.5 \text{ g/Kg}}{14.7 \text{ g/Kg}} = 3\% \quad (2.35)$$

Low levels of humidity have direct results on the quality of life indoors. Plants which exist indoors become dry due to the fast evaporation of the humidity from the soil. Furthermore, human skin becomes dry, there are effects on the nose and the larynx, and bacteria can be more easily introduced into the body.

High values of relative humidity on warm days during summer result in respiratory problems for sensitive members of the population, especially older people and children. When the temperature is high, the main path for lowering of body temperature is through sweating. In the case of low air humidity, evaporation of sweat from the skin occurs quickly and the temperature feels lower than its actual value. In the case of high air humidity, sweating is difficult and the body cannot easily reduce its temperature.

### 2.7.2 Dew Point

The dew point denotes the temperature at which the air has to be cooled without changes of pressure and humidity in order to have saturation. The dew point is an important parameter which is used for the prognosis of ice and fog. At ground level there is no considerable variation of atmospheric pressure and as a result the dew point corresponds to the humidity quantity in air. High values of the dew point indicate a high humidity content and low values of the dew point indicate low humidity content.

The difference between the air temperature and the dew point indicates if the relative humidity is high or low. When the air temperature and the dew point are equal the relative humidity is 100%. This occurs during a snow storm. But the air in a desert area, where there is a large difference between the air temperature and the dew point, has quite low relative humidity. It is interesting to note that the desert air with a higher dew point contains more water vapour than the air in a snowstorm.

There are different phenomena, other than the dew phenomenon, which are related directly with the water vapour condensation. An example is fog formation. When water vapour condenses, visibility is reduced due to growth in size of the atmospheric particles. The size of the fog droplets further increases when the relative humidity is high and the droplets become visible. When the visibility is reduced to less than 1 km, a cloud appears close to the Earth's surface and this is called fog.

The air above urban areas is usually polluted with elevated concentrations of airborne particles. Therefore the fog above urban areas is denser than above the sea under the same atmospheric conditions. Examples include the dense fog which was formed above London in the 1950s. The fog was so dense that the Sun's rays could not penetrate the air and it was necessary to use lamps during the day. The fog can be acid when there are interactions with gaseous pollutants such as sulphur and nitrogen oxides. Furthermore, acid fog has negative consequences to the public health and especially to people with breathing problems.

Fog is formed during evaporation processes, when cold air comes into contact with warm quantities of water vapour. The fog consists of small droplets which are produced from the water vapour condensation at layers of air close to the Earth's surface. A fog is therefore a cloud which is formed in stable layers of air and its base is usually the Earth's surface.

Necessary conditions for the occurrence of fog are:

- small difference between temperature and dew point (large relative humidity)
- presence of condensation nuclei
- weak surface winds
- cold Earth's surface with warm and humid air above or warm sea and cold air above.

Usually the fog is formed in coastal areas where the humidity is elevated and also in industrial areas where there is increased concentration of airborne particles which serve as condensation nuclei.

The occurrence of fog is a result of two main phenomena:

### **2.7.2.1 The Evaporation of Water at Cold Air**

One phenomenon is called evaporation fog and occurs close to the sea surface (or lake), when the air temperature is very low and there is a large temperature difference between sea and air. Its formation is due to the quick evaporation of the sea (or lake) water, and its surface resembles the view of a large boiler that emits large quantities of vapor which are condensed quickly inside the cold air. The layer of fog is small and seldom reaches 30 m and also the visibility changes are highly variable.

### **2.7.2.2 The Freezing of Humid Air**

The fog which is formed with freezing of humid air can be divided into different categories such as radiation fog, advection fog, mixing fog and sea smoke. Different

mechanisms are responsible for these effects. For example the radiation fog is formed during the night without clouds, with a light wind, when a thin layer of humid air is located close to the Earth's surface and under a layer of dry air. The thin layer of humid air does not absorb adequate infrared radiation from the Earth's surface and freezes quickly, while and at the same time it freezes the dry air above. When the temperature equals the dew point, then the condensation of water vapor starts and consequently also the formation of droplets and the fog formation. The interested reader can consult the book by Ahrens (1994) to study the different forms of fog.

### 2.7.3 *Clouds in the Atmosphere*

We explained previously that adiabatic processes together with ascending warm air from the surface results in its expansion and cooling. The unsaturated air mass is cooled adiabatically with a rate close to  $1^{\circ}\text{C}/100\text{ m}$ . When the air is colder, it contains smaller quantities of water vapour. Therefore the air mass which ascends and becomes cooler has also an increase of its relative humidity. At the height at which the temperature of the air mass is equal to the dew point (when the relative humidity reaches 100%) the excessive water vapour starts to condense with the formation of small droplets. The formed droplets coagulate and forms larger droplets. Note that an average rain droplet consists of about  $10^3$  water droplets. Finally a very extensive number of water droplets form a cloud.

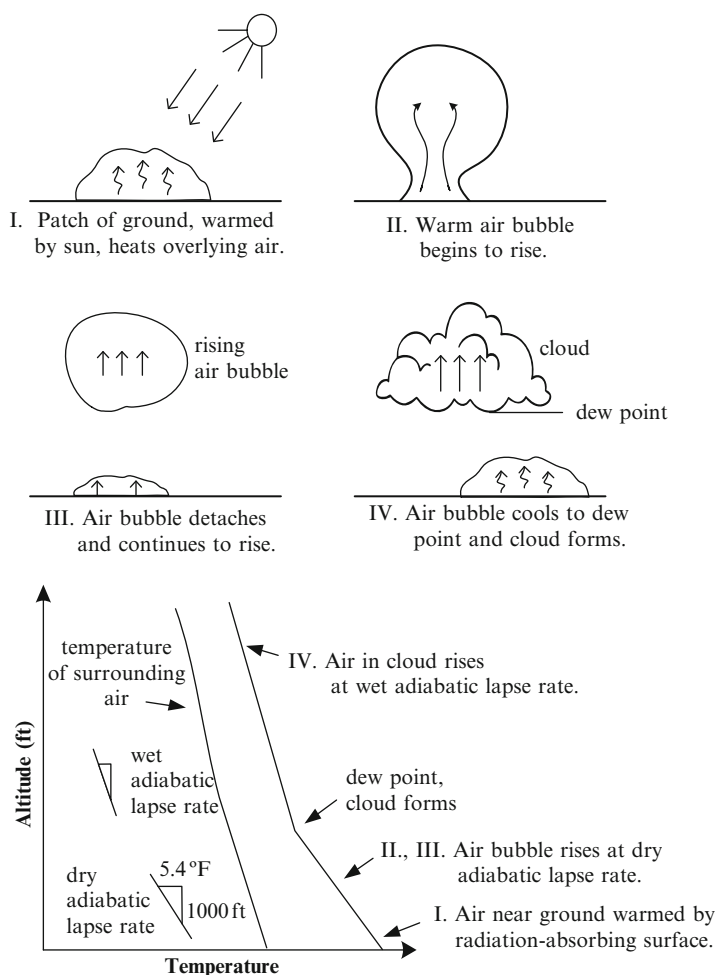
The cloud formation occurs with: (1) radiation of heat from an air mass to the environment, (2) with the air transport to a cooler region and (3) adiabatic rise of an air mass. Figure 2.25 shows the steps in the formation of clouds during adiabatic cooling of warm air masses from the Earth's surface.

As the saturated air mass continuous to rise it will be further cooled with increase of the quantity of water droplets and the cloud size. The temperature lapse rate is not equal to the dry lapse rate since the air mass receives the latent heat which is released from the water vapour during the condensation process. Therefore the temperature drop of the ascending saturated mass is close to  $0.5^{\circ}\text{C}/100\text{ m}$  which is known as wet lapse rate.

It is accepted that most clouds are formed with adiabatic cooling which occurs inside air masses when they ascend inside the atmosphere. The ascending movement of air masses is due to:

- The vertical air transport after intense surface heating,
- The impact of air masses on mountains,
- The convergence of air masses due to barometric systems,
- The movement of air at the warm and cold fronts.

For the formation of clouds it is not enough to have only adiabatic cooling but it is necessary that condensation nuclei exist in the atmosphere. The water vapour condensation can occur on the surface of soluble (such as NaCl) or insoluble particles (such as dust or diesel particles) and on ions in the atmosphere.



**Fig. 2.25** Schematic representation of cloud formation in the atmosphere with adiabatic cooling of a warm air volume from the Earth's surface (Adapted from Hemond and Fechner 2000)

### 2.7.4 Precipitation

The term precipitation refers to all condensation products of water that fall from the atmosphere as rain, snow, hail and soft hail. The formation of precipitation inside clouds is determined from the air temperature and the turbulence conditions in the atmosphere. Precipitation is one of the most important meteorological and climatological parameters. During precipitation a factor which is important to be studied is the water quantity that drops on a surface and is referred to as rainwater. This

quantity expresses the rainwater height (more commonly referred to as depth) on a horizontal surface and is measured with a rain-gauge. Another useful parameter in climatology is the rain intensity which expresses the resulting rainwater height per unit time. Internationally the measurement unit of rainwater height is mm or cm. For example 1 mm rainwater expresses a water quantity of 1 kg water per 1 m<sup>2</sup> surface.

In respect to the droplet size and precipitation conditions, rain has different names such as “shower” which is produced from clouds of vertical development and has sudden starts and stops as well as abrupt changes in intensity. Drizzle is characterized by small and many droplets which are suspended and follow the air currents.

For the study of the water precipitation in a region it is necessary to consider the following parameters:

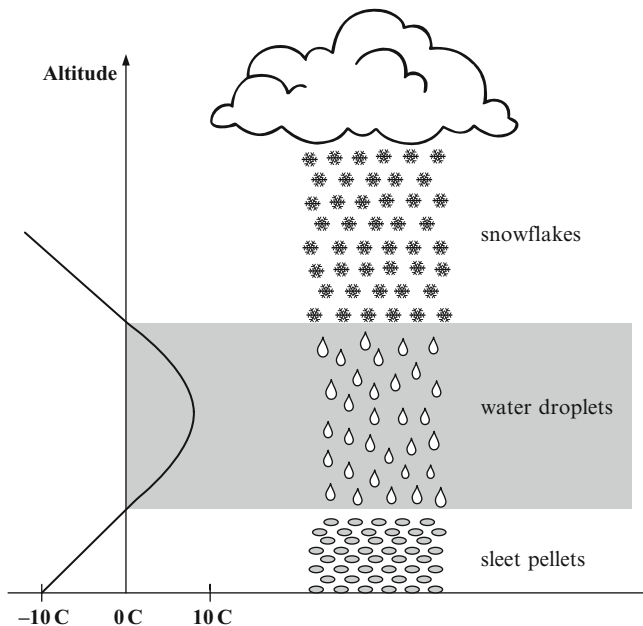
- **Average Monthly Precipitation (A.M.P.):** The average total water precipitation per month (T.P.). For example the average total water precipitation in January in the period 1959–2004 is equal to the summation of the total January precipitation per year divided by the total number of observation years:

$$\text{A.M.P. (JAN)} = \frac{T.P.(\text{JAN}1959) + T.P.(\text{JAN}1960) + \dots + T.P.(\text{JAN}2004)}{46} \quad (2.36)$$

- **Total maximum precipitation of 24 h (T.max.P.24 h):** The maximum water precipitation during 24 h. The maximum water precipitation is observed and logged during 24 h in a period of 1 month for the total number of observation years. This value is possible to be characterized as extreme and it is important to know the year that it occurred. The T.max.P.24 h has to be considered for the occurrence of floods.
- **Average Precipitation Days (A.P.D.):** The average number of days in a month at which there is water precipitation. The minimum water amount which is necessary to precipitate in order to be considered as water precipitation is different among countries but are generally close to 0.1 mm. For example for the calculation of the average precipitation days of January in the period 1959–2004 are equal to the summation of the precipitation days per year divided by the total number of observation years:

$$\text{A.P.D.} = \frac{P.D.(\text{JAN}1959) + P.D.(\text{JAN}1960) + \dots + P.D.(\text{JAN}2004)}{46} \quad (2.37)$$

The atmosphere is a dynamic system and changes in the form of precipitation can occur during transport from its origin to the Earth’s surface. Figure 2.26 shows some physical processes which can be associated with precipitation.



**Fig. 2.26** Snowflakes originating from a cloud can melt during their downward movement when they encounter a warm layer of air and after can freeze again and form sleet pellets when they meet a colder air layer

### 2.7.5 Study of Precipitation Scavenging

There are several ways for precipitation to occur as discussed in the current chapter. The most common ones are through rainfall and snowfall. The flux of gasses and particles from the atmosphere to the Earth's surface through rain can be defined as (Seinfeld and Pandis 2006)

$$W_{gas/rain}^i = \Lambda_{ig} C_{i,gas}, \quad (2.38)$$

$$W_{aeros/rain}^i = \Lambda_{ip} C_{i,part}, \quad (2.39)$$

where  $\Lambda_{ig}$  and  $\Lambda_{ip}$  are the scavenging coefficients for the components  $i$  for the gaseous and particulate phase respectively. The total scavenging  $F_{bc}(t)$  ( $\text{Kg m}^{-2} \text{h}^{-1}$ ) under clouds, when the concentration of the pollutants exists in a horizontally homogeneous atmosphere, is  $C_g(z,t)$ ; then it results that

$$F_{bc}(t) = \int_0^h \Lambda_g(z,t) C_g(z,t) dz, \quad (2.40)$$

where  $h$  is the height of the cloud base and  $\Lambda_g$  is the scavenging coefficient which is dependent on time ( $s^{-1}$ ). The total scavenging is the sum of the scavenging inside the clouds (washout) and under the clouds (rainout). For a homogeneous atmosphere under the clouds it can be written that

$$F_{bc}(t) = C_g(t) \int_0^h \Lambda_g(z, t) dz = \bar{\Lambda}_g h C_g(t), \quad (2.41)$$

where  $\bar{\Lambda}_g$  is the average value of the scavenging coefficient.

The washout ratio can be defined as

$$w_r = \frac{C_{i,precip}(x, y, 0, t)}{C_{i,air}(x, y, 0, t)}, \quad (2.42)$$

where  $C_{i,precip}(x, y, 0, t)$  is the concentration of component  $i$  which is contained inside the rain at the Earth's surface and  $C_{i,air}(x, y, 0, t)$  is the concentration of component  $i$  inside the air at the Earth's surface. Therefore the flux  $F_w$  of the wet scavenging can be expressed as (Seinfeld and Pandis 2006)

$$F_w = C_{i,precip}(x, y, 0, t) p_o, \quad (2.43)$$

where  $p_o$  is the rain intensity ( $mm\ h^{-1}$ ). For light rain the rain intensity is equal to  $p_o = 0.5\ mm\ h^{-1}$ , whereas for heavy rain  $p_o = 25\ mm\ h^{-1}$ . Furthermore, the velocity of wet deposition can be defined as  $u_w = \frac{F_w}{C_{i,air}(x, y, 0, t)}$ .

In meteorology the rain is defined when the descending droplets have a diameter larger than or equal to 0.5 mm. After an intense rain there is usually better visibility since rain is scavenging a large number of particles. The rain has also the ability to absorb water soluble chemical components and remove them from the atmosphere. It is important to calculate the scavenging rate of gaseous components in the atmosphere based on knowledge of the rain's characteristics such as density and droplet size, as well as from the physico-chemical characteristics of the gaseous species.

The transfer of gasses on the droplet surface can be calculated from the expression

$$W_t(z, t) = K_c (C_g(z, t) - C_{eq}(z, t)), \quad (2.44)$$

where  $K_c$  is the coefficient of mass transfer ( $cm\ s^{-1}$ ),  $C_g$  is the concentration of the pollutant under study in the gaseous phase and  $C_{eq}$  its concentration on the droplet surface which is in equilibrium with the concentration in the aqueous phase.

Using Henry's law we can derive that

$$C_{eq} = \left( \frac{1}{H} \right) C_{aq}, \quad (2.45)$$

where,  $H$  is Henry's coefficient and  $C_{aq}$  is the pollutant concentration in the aqueous phase.

Consequently the equation (2.44) can be written as

$$W_t(z, t) = K_c \left( C_g(z, t) - \frac{C_{aq}(z, t)}{H} \right). \quad (2.46)$$

The coefficient of mass transfer from molecules in the gaseous phase to the droplet can be calculated from the expression (Seinfeld and Pandis 2006)

$$K_c = \frac{D_g}{D_p} \left[ 2 + 0.6 \left( \frac{\rho_{air} U_t D_p}{\mu_{air}} \right)^{1/2} \left( \frac{\mu_{air}}{\rho_{air} D_g} \right)^{1/3} \right], \quad (2.47)$$

where,  $D_p$  is the droplet diameter,  $D_g$  is the diffusivity in the gaseous phase,  $\rho_{air}$  is the air density,  $\mu_{air}$  is air viscosity and  $U_t$  is the droplet velocity. In the above equation the term  $Sh = K_c D_p / D_g$  is the Sherwood number,  $Re = \rho_{air} U_t D_p / \mu_{air}$  is the Reynolds number and  $Sc = \mu_{air} / \rho_{air} D_p$  is the Schmidt number. The concentrations  $C_{aq}$  and  $C_g$  are functions of height and time.

In addition to precipitation occurring under clouds, transport of chemical components to rain droplets can occur inside clouds. Gaseous species such as  $HNO_3$ ,  $NH_3$ , and  $SO_2$  can be absorbed inside rain droplets. If the concentration of particles inside a cloud is equal to  $N(D_p)$  then the transfer rate of a gas  $W_{ic}$  which has larger concentration in air than in the liquid phase (e.g.,  $HNO_3$ ) is given by the expression (Seinfeld and Pandis 2006)

$$W_{ic} = C_g \int_0^\infty K_c \pi D_p^2 N(D_p) dD_p = \Lambda C_g, \quad (2.48)$$

where  $\Lambda$  is the scavenging rate. A typical concentration of droplets can be given by the expression  $N(D_p) = \alpha e^{-bD_p}$  where  $\alpha = 2.87 \text{ cm}^{-4}$ ,  $b = 2.65 \text{ cm}^{-1}$  at  $D_p = 5\text{--}40 \text{ }\mu\text{m}$ . With a replacement in equation (2.48) it can be concluded that  $\Lambda = 0.2 \text{ s}^{-1}$  for a cloud which consists of  $288 \text{ droplets/cm}^3$ . The above calculations denote that the processes which transfer chemical species at droplets are very fast, of the order of a few seconds in comparison with the transport of the chemical species inside a cloud or with changes of the ratio of condensation and evaporation. Several chemical species, such as  $SO_2$  are not absorbed directly from rain droplets. Their absorption is dependent on other factors such as the presence of other chemical species inside droplets and the pH of the water. In addition to the absorption of gaseous species there is also absorption of particles inside clouds. A common process is the increase of particle size through absorption of water vapour and coagulation with other particles.

## 2.8 Applications and Examples

**Example 1.** An air volume is rising adiabatically from height  $z_1$  to  $z_2$ . Prove that the relationship between temperature and pressure at two different heights in the atmosphere are  $\frac{T(z_2)}{T(z_1)} = \left[ \frac{p(z_2)}{p(z_1)} \right]^{(\gamma-1)/\gamma}$ .

From the theory it is known that

$$\frac{dT}{dp} = \frac{m R T / M_{air} p}{C_v + m R / M_{air}},$$

where

$$\hat{c}_p - \hat{c}_v = \frac{R}{M_{air}}$$

And

$$\gamma = \frac{\hat{c}_p}{\hat{c}_v}.$$

Therefore:

$$\begin{aligned} \frac{dT}{T} &= \frac{mR}{M_{air}} / \left( c_v + \frac{mR}{M_{air}} \right) \frac{dp}{p} \Rightarrow \frac{dT}{T} = \frac{R}{M_{air}} / \left( \hat{c}_v + \frac{R}{M_{air}} \right) \frac{dp}{p} \\ \Rightarrow \int_{T(z_1)}^{T(z_2)} \frac{dT}{T} &= \int_{p(z_1)}^{p(z_2)} \left( \frac{\gamma - 1}{\gamma} \right) \frac{dp}{p} \Rightarrow \ln \frac{T(z_2)}{T(z_1)} = \left( \frac{\gamma - 1}{\gamma} \right) \ln \left( \frac{p(z_2)}{p(z_1)} \right). \end{aligned}$$

Finally:  $\frac{T(z_2)}{T(z_1)} = \left[ \frac{p(z_2)}{p(z_1)} \right]^{(\gamma-1)/\gamma}$ .

If  $z_1$  is the height of the Earth's surface and air volume, which is initially at conditions  $T, p$  is rising adiabatically to pressure  $p_o$ , then the temperature  $\theta$  at pressure  $p_o$  is given by the relation  $\theta = T \left( \frac{p}{p_o} \right)^{-\frac{(\gamma-1)}{\gamma}}$  and is called potential temperature

**Example 2.** Show that if the atmosphere is isothermal, then the temperature change of the temperature of an air volume which, ascending adiabatically, is given by the expression  $T(z) = T_o e^{-\Gamma z / T_o}$  where  $T_o$  and  $T_o'$  are the temperatures of the air volume and the atmospheric air at the surface respectively.  $\Gamma = \frac{g}{\hat{c}_p}$ .

For the air volume ( $T'$  is the air temperature and  $T$  is the temperature of the air volume under study) we have

$$\frac{dp}{dz} = - \frac{M_{air} g p}{R T'}.$$

For the air volume which is ascending we can write that

$$\frac{dT}{dp} = \frac{m R T / M_{air} p}{C_u + m R / M_{air}} \Rightarrow \frac{dT}{dp} = -m g \frac{\frac{dz}{dp} \frac{T}{T}}{c_v + m R / M_{air}}$$

which results in  $\frac{dT}{dz} = -\Gamma \frac{T}{T}$ .

With the integration of the above expression it follows that

$$\int_{T_o}^T \frac{dT}{T} = \int_0^z -\Gamma \frac{dz}{T} \Rightarrow \ln \frac{T}{T_o} = -\frac{\Gamma}{T_o} z \Rightarrow T(z) = T_o e^{\left(-\frac{\Gamma z}{T_o}\right)}.$$

**Example 3.** Concentrations of two gaseous species ( $N_1 = 8$  και  $N_2 = 4$ ) with velocities ( $u_1 = 3$  and  $u_2 = -1$ ) have been measured at two different points at specific times. Calculate the following variables:  $\bar{N}$ ,  $N'_1$ ,  $N'_2$ ,  $\bar{u}$ ,  $u'_1$ ,  $u'_2$ ,  $\overline{u'N'}$ ,  $\bar{u}\bar{N}$ ,  $\overline{uN}$ .

Using the expressions for the average and instant values it can be written that

$$\bar{N} = (N_1 + N_2)/2 = 6,$$

$$N'_1 = N_1 - \bar{N} = 2,$$

$$N'_2 = N_2 - \bar{N} = -2,$$

$$\overline{u'N'} = (u'_1 N'_1 + u'_2 N'_2)/2 = 4,$$

$$\overline{uN} = \bar{u}\bar{N} + \overline{u'N'} = (u_1 N_1 + u_2 N_2)/2 = 10 :$$

$$\bar{u} = (u_1 + u_2)/2 = 1,$$

$$u'_1 = u_1 - \bar{u} = 2,$$

$$u'_2 = u_2 - \bar{u} = -2,$$

$$\bar{u}\bar{N} = 6.$$

**Example 4.** A meteorological station measures with anemometers the components  $U$  and  $W$  of the wind. Velocity measurements have been performed every 6 s for 1 min. The measurements are given in the following 10 pairs:

$U$  (m/s) 5 6 5 4 7 5 3 5 4 6,  
 $W$  (m/s) 0 -1 1 0 -2 1 2 -1 1 -1.

Calculate the average value and dispersion of each component of the velocity, as well as the correlation coefficient between  $U$  and  $W$ .

Using the expressions for the average and correlation coefficient values, the following values can be written.

$$\overline{U} = 5 \text{ m s}^{-1}$$

$$\overline{W} = 0 \text{ m s}^{-1}$$

$$\sigma_U^2 = 1.20 \text{ m}^2 \text{ s}^{-2}$$

$$\sigma_U = 1.10 \text{ m s}^{-1}$$

$$\sigma_W^2 = 1.40 \text{ m}^2 \text{ s}^{-2}$$

$$\sigma_W = 1.18 \text{ m s}^{-1}$$

$$\overline{u' w'} = -1.10 \text{ m}^2 \text{ s}^{-2}$$

$$r_{UW} = -0.85$$

From the above expressions can be concluded that the turbulence component of velocity  $W$  is higher than the  $U$  component even if the average value for the velocity  $W$  is zero. From the negative value of the correlation coefficient  $r_{uw}$  it can be concluded that the variations of the components  $U$  and  $W$  are occurring mainly in opposite directions.

**Example 5.** The concentration of a gas  $A$  is equal to  $10 \mu\text{g m}^{-3}$  under a cloud. Suppose that the washout coefficient is constant and equal to  $3.3 \text{ h}^{-1}$ . Calculate the concentration of gas  $A$  in the atmosphere after 30 min of rain and the total flux of the wet deposition. The cloud base is 2 km (Adapted from Seinfeld and Pandis, 2006).

The concentration variability can be expressed as

$$\frac{\partial C}{\partial t} = -W_{\text{air/rain}} + R + E,$$

where since there no emissions or chemical reactions ( $R = 0$ ,  $E = 0$ ) it can result that

$$\frac{\partial C}{\partial t} = -\Lambda C,$$

where  $\Lambda$  is the washout coefficient. The solution of the equation gives

$$C = C_o e^{-\Lambda t}.$$

After 30 min the resulting concentration is equal to  $C = C_o \times 0.19 = 1.9 \mu\text{g m}^{-3}$ . Therefore,  $C_o - C = 8.1 \mu\text{g m}^{-3}$  and the total flux for a column of 2 km is equal to  $8.1 \times 2,000 = 16.2 \text{ mg m}^{-2}$ .

### Problems

**2.1** It is proposed that the problem of air pollution in the city of Los Angeles can be solved by digging tunnels in the surrounding mountains and pumping the air outside to the surrounding areas which are mainly deserts. Calculate the energy that would be needed for the transport of air from Los Angeles. Los Angeles covers an area of  $4,000 \text{ km}^2$  and the polluted air is located under the boundary layer which has an average height of 400 m. The viscosity coefficient of air which is transported above the Los Angeles area is 0.5 and the minimum energy which is required for the air flux is equal to the energy which is consumed from the surface friction. Calculate the energy which would be required for an air mass transport with velocity 7 km/h. Compare the result with the capacity of Hoover Dam (power station of energy production) which is equal to  $1.25 \times 10^6 \text{ KWh}$ .

**2.2** The value of the vertical temperature lapse rate in an area is equal to  $5^\circ\text{C/km}$  and the temperature of air on the surface is equal to  $20^\circ\text{C}$ . If an insulated balloon full of dry air with temperature  $50^\circ\text{C}$  is allowed to ascend from the surface, then calculate the height which the balloon can reach (the balloon temperature lapse rate is equal to  $\gamma_d = 10^\circ\text{C/km}$ ).

**2.3** Simultaneous measurements of the air temperature at four points A, B,  $\Gamma$  and  $\Delta$  which are located downwind of a mountain chain are the following:

Locations	A	B	$\Gamma$	$\Delta$
Height (m)	1,530	1,396	690	378
Temperature (T) ( $^\circ\text{C}$ )	6.3	6.9	11.9	14.4

Calculate the value of the vertical lapse rate between the positions (a)  $\Delta \Gamma$ , (b)  $\Gamma B$ , (c)  $B A$  and (d)  $\Delta A$ . Additionally make a comparison of the mean value of the vertical lapse rate above the mountain chain with the corresponding value of  $-6.5^\circ\text{C/km}$  and determine the difference.

**2.4** A potential temperature  $\Theta$  is determined under the hypothesis of adiabatic processes. Changes in entropy can be connected with changes in temperature and pressure with the expression

$$dS = \left( \frac{\partial S}{\partial T} \right)_p dT + \left( \frac{\partial S}{\partial p} \right)_T dp.$$

a. Show that

$$dS = \left(\frac{c_p}{T}\right)dT - \left(\frac{R}{M_a}\right)\frac{dp}{p}.$$

b. During adiabatic conditions ( $dS = 0$ ) show that  $d\Theta = 0$  and  $\Theta = \text{const} \tan t \times \frac{T}{p^{(\gamma-1)/\gamma}}.$

**2.5** In an urban area, the air temperature at 07:00 after a cloudless night in January is equal to  $0^\circ\text{C}$ . The height of a temperature inversion was  $h = 0$  and its depth  $d = 100$  m. Calculate at which height from the inversion base (surface) it will be the same air temperature, if the value of the vertical temperature lapse rate is  $-(\theta T/\theta z) = 5^\circ\text{C}/\text{km}$  and the intensity of the inversion is equal to  $0.1^\circ\text{C}/10\text{m}$ .

**2.6** Calculate the concentration ( $\text{mole}/\text{cm}^3$ ) and the mixing ratio (ppm) of water vapour at the Earth's surface for temperature equal to  $T = 298\text{K}$  and relative humidity RH 50%, 60%, 70%, 80%, 90%, 95%, 99%. The water vapour pressure of pure water (saturation pressure) versus temperature is given by the expression

$$p_{\text{H}_2\text{O}}^0(T) = p_s \times \exp[13.3185\alpha - 1.976\alpha^2 - 0.6445\alpha^3 - 0.1299\alpha^4],$$

where:

$$p_s = 1013.25\text{mbar}$$

and

$$\alpha = 1 - \left(\frac{373.15}{T}\right).$$

**2.7** The prognosis of the daily atmospheric temperature structure under non-variable conditions is examined here. In this case it is necessary to examine the spatial temperature changes only in the vertical direction. It is assumed that the radiation absorption from the atmosphere is negligible and the dynamic temperature  $\theta$  can be determined from the expression

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left( K \frac{\partial \theta}{\partial z} \right). \quad (2.49)$$

In the above expression it is assumed that  $K$  has a constant value. It is also assumed that at elevated heights the temperature profile is the same with an adiabatic rate equal to

$$\theta \rightarrow 0 \quad \text{when} \quad z \rightarrow \infty. \quad (2.50)$$

The temperature at the surface ( $z = 0$ ) is dependent on the Sun's heat during the day and the cooling at night is due to radiation emission. Therefore  $\theta(0,t)$  can be expressed as

$$\theta(0,t) = A \cos \omega t. \quad (2.51)$$

where  $A$  is the width of the daily variation and  $\omega = 7.29 \times 10^{-5} \text{ s}^{-1}$ .

A) Show that the solution which satisfies the equations (1 – 3) is

$$\theta(z,t) = A e^{-\beta z} \cos(\omega t - \beta z),$$

where  $\beta = \sqrt{\omega/2K}$ . The above solution is called a steady-state solution and describes the temperature dynamics which correspond to the influence of equation 3.

B) Show that the height  $H$  which expresses the base or the top of a temperature inversion is given by the expression

$$\sin(\omega t - \beta H) - \cos(\omega t - \beta H) = \frac{g}{A\beta\hat{c}_p} e^{\beta H}.$$

Is it possible to have more than one temperature inversion ?

Show that  $\frac{\partial \theta}{\partial z} = \frac{\partial T}{\partial z} + \Gamma$

where  $\Gamma = \frac{g}{c_p}$ .

**2.8** It is observed in an area that the atmospheric temperature is decreased by 14 K between the Earth's surface and height of 2 km above it. What is the vertical lapse rate and how does it compare with the dry and wet lapse rate ? Calculate the atmospheric stability conditions under the above conditions.

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