Vlado Spiridonov
Mladjen Ćurić

Fundamentals of Meteorology

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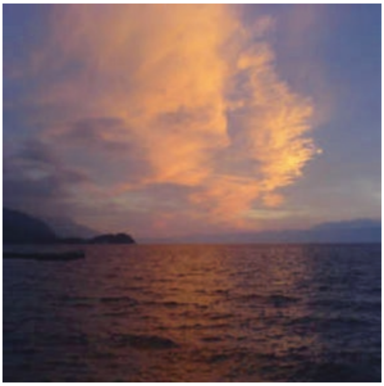
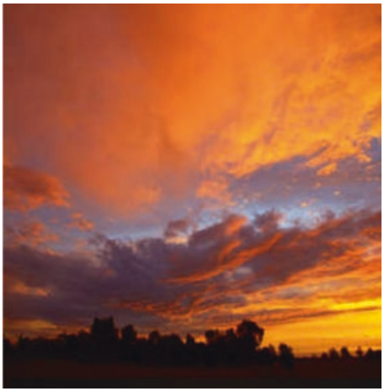
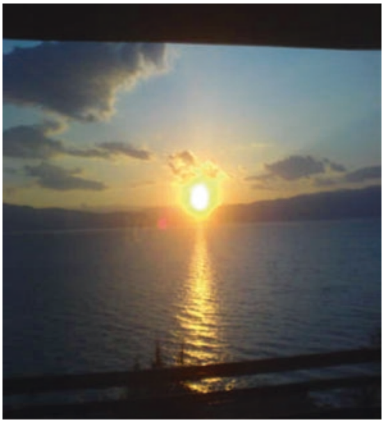
Preface







This book is dedicated to our planet's atmosphere, our great challenge and inspiration. It is just an extended arm of our great fascination, "Meteorology." The book opens the doors to the achievements made in this new millennium, making linkage between contemporary science, technology and society, and the pathways how meteorology contributes to the development of science. *Fundamentals in Meteorology* incorporates all significant eras, from ancient times to the present days. It explores important atmospheric phenomena and physical processes from a local to global scale and from seconds to years. In addition to the general topics, our novel book incorporates other important ones such as the atmospheric boundary layer, atmospheric waves, atmospheric chemistry, optics, and electricity.

As the weather keeps changing time to time and from place to place, sudden worsening of its conditions can change our mood, especially when it is unfavorable, unstable, and severe. But when the sun and the blue sky reappear, the rays of happiness return our confidence, which is a sufficient reward for this modest work. Finally, we invite you to join us on a journey through the atmospheric space and bear witness to these wonderful events occuring within our atmosphere and even beyond.

Wien, Austria — Vlado Spiridonov
Belgrade, Serbia — Mladjen Ćurić

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Finally, we would like to thank our families for their unreserved support, patience, and understanding throughout the years for this hard and delicate work such as writing a book.

Thank you.
By authors

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About the Authors

Mladjen Ćurić was born in Zabljak, Montenegro. He completed his Bachelor in Hydrometeorology l in Belgrade and was awarded top marks. Dr. Ćurić graduated in meteorology from the University of Belgrade, where he also completed his post-graduate study and Ph.D. thesis. From 1977 to 1980, he attended Colorado State University, Fort Collins, USA, through several study visits; Imperial College, London, England; Manchester University, England; and The Meteorological Office, Bracknell. From 1972 to 1978, Dr. Ćurić was appointed as Teaching Assistant, Department of Physics and Meteorology, Belgrade University, and since 1990 he has been a Full Professor at the Institute of Meteorology, Belgrade University. From 1979 to 1982, Dr. Ćurić was Director of the Institute of Meteorology, University of Belgrade. From 1982 to 1984, he was Vice Dean in the Faculty of Physics, University of Belgrade. Dr. Ćurić was also nominated as a Member of the Executive Committee of the International Commission on Clouds and Precipitation-IAMAP from 1988–1996. Since 1990, he has been a Member of the Executive Committee of the National Association of Environment Protection. Dr. Ćurić was Vice Dean at the Faculty of Physics from 1996 to 1998 and from 2004 to 2007.

His professional and scientific interest are the fundamental topics in meteorology, that is, dynamics of atmosphere, cloud physics, applied meteorology and hydrology, weather modification, and environment protection. Dr. Ćurić is the author of several books and author and co-author of more than two hundreds papers, mainly in international reviewed journals (*Journal of the Atmospheric Science*, *Quarterly Journal of the Royal Meteorological Society*, *Tellus*, *Journal of Applied Meteorology*, *Atmosphere-Ocean*, *Theoretical and Applied Climatology*, *Atmospheric Research*, *Meteorology and Atmospheric Physics*, *Journal of Geophysical Research*).

Vlado Spiridonov was born in Skopje. He completed primary school in Grigor Prlicev, secondary school in Rade Jovcevski Korcagin, and high school at the Institute of Physics, the Faculty of Natural Sciences, St. Cyril and Methodius University, Skopje. Dr. Spiridonov received specializing Sci., M.Sc., and Ph.D. degrees from the Institute of Meteorology, Physical Faculty of the University of Belgrade.

In 2004, the National Research Council of Canada granted a Postdoctoral Fellowship to Dr. Spiridonov at Meteorological Service of Canada, The Air Quality Modelling Branch.

In the same year, the Government of the Republic of Macedonia appointed Dr. Spiridonov as Director General of the Hydrometeorological Service and the Permanent Representative of the Republic of Macedonia to the World Meteorological Organization.

In April 2007, he was appointed as Assistant Professor in Meteorology at the Faculty of Natural Sciences and Mathematics, while in 2012 he became Associate Professor in Meteorology, and still continuing, at Ss Cyril and Methodius University in Skopje. Dr. Spiridonov is a Faculty Member in the National Council of Climate Change in Macedonia as well as the Representative of the Commission of Atmospheric Science at the World Meteorological Organization.

From 2017 to 2019 Dr. Spiridonov was a Visiting Professor of Meteorology at the University of Vienna, Faculty of Earth Sciences, Geography and Astronomy, Department of Meteorology and Geophysics, as a proxy for General and Theoretical Meteorology. He has made a significant contribution in developing and promoting education and learning platforms by implementation of innovative approach and novel solutions in academic practice at UNIVIE. His research activities and valuable scientific publication record within a 2 years were highly acknowledged by the Department of Meteorology and Geophysics.

He has participated in several international scientific conferences, organized international events and symposiums, and has given presentations, scientific lectures, and invited talks to plenary sessions and seminars worldwide.

Dr. Spiridonov has published many scientific papers in international journals, five books, and has won several awards (e.g., Innovation of the Year-2000 in Macedonia, Gold medal with mention at EUREKA-2000 exhibition in Brussels, a Genius Prize by Hungarian Association of Innovations, and GRAND PRIX in 2002 on 4th GENIOUS International Invention Exhibition in Budapest for the invention associated with fog dispersal method).

His present research is related to the development of a new storm alert system using a non-hydrostatic mesoscale forecast model, a Cloud Resolving Model, and suitable diagnostic tool for calculation of severe weather indices. Many sensitivity studies indicate that this novel method shows good performance in more accurate evaluation of severe storms and the local-scale hazards impact over area of interest as valuable information for severe weather alert worldwide.

Chapter 1
Introduction

Meteorology belongs among the oldest scientific disciplines. It has a long-lasting history that reaches back to the distant past of the human civilization. From the beginnings of life, people have been trying to adapt to the atmospheric behaviour, to create a more comfortable life, and to control those conditions that endanger or make life uncomfortable. There is no doubt that today meteorological science has a tremendous importance around the world. Every day, people are faced with various manifestations of weather. Depending on weather conditions, they adapt their planned activities. Weather manifestation is closely related to the status of the atmosphere, which has a wave nature. The scientific discipline that examines atmospheric phenomena, atmospheric structure, composition, atmospheric features and phenomena, and the important processes that occur in a thin surrounding layer of the atmosphere-troposphere, water, and air including the future state of the atmosphere is referred to us as *"meteorology"*. Basically, meteorology examines the physical processes that occur in the atmosphere. Hence, this scientific discipline is often referred to as the *"Physics of the Atmosphere"* (see Spiridonov 2010; Spiridonov and Ćurić 2011; Andrews 2010; Salby 1996). Meteorology as an interdisciplinary science cuts across a various number of natural disciplines. Thus, there is a great interest in meteorology, especially in the last two to three decades, when the global community is facing with global warming and climate change with extreme weather events. During this period many comprehensive textbooks, scientific publications, and studies were dedicated to this contemporary and interdisciplinary science (e.g. Andrews 2010; Ackerman and Knox 2007; Ćurić 2006; Lutgens and Tarbuck 2009; Salby 1996; Anthes 1996).

The key motivation for writing this contemporary textbook arrives from the need for comprehensive and systematic way of describing all the important atmospheric phenomena and processes that constitute the modern concept of meteorology. The purpose is to offer an advanced understanding and significant knowledge of the important topics relating to these natural environmental processes. Standard, theoretical, and experimental approach gradually adapts the reader into the subject and introduces the problems which emerge. Furthermore, the essential elements of

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mathematics are used, in order to easily present certain phenomena and processes in the atmosphere. We have tried to incorporate all important topics in the modern concept of the physics of atmosphere meteorology. The writing is primarily intended for students of the atmospheric and environmental sciences and physics of the atmosphere meteorology or those who attend the general course in meteorology. Yet again, this book is designed for a wider circle of readers, starting from primary, secondary, and higher education to all those interested in meteorology who want to find useful information, content, and interpretation of certain global phenomena and processes occurring in the atmosphere of our planet. With an exceptionally readable, comprehensive, and extensive illustrative and interesting approach with the standard appropriate mathematical concept, it describes the basic characteristics of the atmosphere, including weather, climate, and climate change. The book begins with the definition of the subject and the tasks of meteorology, methods of research, classification of meteorology, and the relation between meteorology and other sciences. It also contains chapters on the historical overview, structure and chemical composition of the atmosphere, energy and radiation in the atmosphere, the energy budget of the Earth, the basis of the thermodynamics of the atmosphere, air temperature, and its variations. The book covers all the major topics of atmospheric moisture processes (e.g. air humidity, condensation and formation of clouds, their classification as well as an explanation of the process of formation of precipitation). It also deals with chapters devoted to air pressure and winds, atmospheric statics, planetary boundary layer, atmospheric dynamics, turbulence, the global atmospheric circulation, air masses and fronts, cyclones, anticyclones, and tropical cyclones. The book also contains basic information about natural disasters related to weather, water and climate, atmospheric optical phenomena (photometers), and atmospheric electricity. Climate and climate change, complexity of climate system, climate models, the modern watching of climate change, greenhouse effect, and global warming are considered with a special attention. A chapter in this book is especially devoted to the methods and techniques of analysis and weather forecasting development and application of numerical weather prediction. The 25th novel chapter briefly introduces the readers in atmospheric chemistry, aerosols, and the factors affecting the pollution source of atmospheric gases and aerosols. The last chapter describes the meteorological measurements and observations, modern instruments and devices used to measure atmospheric phenomena, radar and satellite measurements, and observations of the atmosphere.

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Chapter 2
Meteorology as a Natural Science

Meteorology is the branch of science which studies the atmosphere of our planet; its structure, composition, and properties; the physical processes closely related to the Earth's surface, water, and air; various weather phenomena; weather and climate; and the future state of the atmosphere (Lindzen et al. 1990; Ackerman and Knox 2007; Ahrens and Henson 2016).

2.1 Definition of the Atmosphere

According to the AMS glossary of meteorology, Earth's atmosphere is a gaseous envelope gravitationally bound to our planet Earth (Fig. 2.1) (Glickman 2000). In order to determine its status and a behaviour at a given time, it is necessary to conduct observations at a different point, not only at the Earth's surface (surface observations) but also at a certain height above sea level (upper air observations). From here comes the necessity to organize a special meteorological station network and aerological observation, equipped with suitable instruments and apparatus. Moreover, there are special observatories for detailed observation and monitoring of certain electrical, optical, and sound phenomena, the turbulence in the atmosphere, and certain physical processes in clouds. Generally, in studying physics, the laboratory experiment appears as the primary used method. On the other hand, the meteorological laboratory is the atmosphere itself. In research applied in physics, conditions can be changed to introduce or eliminate certain factors in order to adopt some laws.

Unlike physics, in meteorology such changes are not possible because numerous secondary factors that affect a phenomenon or processes cannot be isolated. As the scientific discipline, meteorology has the following tasks:

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Fig. 2.1 The Earth's atmosphere

1. To identify and describe the processes that emerge in the atmosphere in a qualitative and quantitative manner
2. To explain these phenomena and based on the obtained data to establish the laws governing these processes
3. By applying these laws, to develop secure and reliable methods to predict their future development in a period
4. To develop effective methods for modifying natural atmospheric conditions based on personal needs, interdisciplinary application of meteorology in all economic sectors (e.g. energy, transport, health, environment, tourism), and other important activities

Meteorology is the contemporary natural science that is applied in various sectors (e.g. water management, industry, energy, agriculture, transport, health, tourism, etc.).

2.2 Methods of Research of the Atmosphere

In the current stage of development, meteorology as a natural science offers four methods of scientific research:

1. Method of observation in natural conditions
2. Method of experiments, which in a wider sense means laboratory measurements and research in natural conditions, so-called field experiments
3. Method of theoretical analysis based on the general laws of physics, the physical principles (Blake and Robson 2008), and exploitation of mathematical apparatus
4. Numerical methods for solving processes and phenomena considered a subject of meteorology (atmospheric models)

A wide application in meteorology has the experimental method of measurement and monitoring, in such cases when a specific data for scientific research purposes (Antarctica, Arctic, oceanographic observations, etc.) are required.

2.2.1 Experimental Method of Research

The experiment has a significant place in meteorology. Certain atmospheric phenomena, such as lightning, cloud formation, and polar lights, could be monitored in the laboratory conditions. In meteorology a huge amount of data that are subjected to processing and analysis could be collected through observation and experiment. In that view, the statistical method has the primary importance. It allows, by way of averaging, to exclude random sides of separate atmospheric phenomena and to separate primary direction with its characteristics. Statistical method of correlation represents statistical measure that describes the size and direction of a relationship between two or more meteorological variables. This method quantitatively expresses the extent of that relationship. However, statistical methods do not explain in more detail these relationships. For these reasons, the ultimate need is to apply an essential mathematical-physical analysis based on physical laws, specifically the physical and mathematical relations for description of the atmospheric behaviour.

2.3 Relationship Between Meteorology and Other Sciences

As a planet, Earth is consist of three substances: gas (atmosphere), liquid (hydrosphere), and solid (lithosphere). Physical-chemical processes that take place within them are now taught by independent studies. All these components are integrated under a common name known as *"geophysics"*. Consequently, meteorology is the group of geophysical sciences, and therefore it is closely related to other sciences of that group, such as Earth physics and oceanography. Since a long time ago, the linkage between meteorology and astrophysics, and more specifically between meteorology and physics of the Sun, is established and reinforced. Regardless of the numerous links to other sciences, according to the nature of task solution and methods used, meteorology is a physical science. It can be successfully developed only based on achievements in physics, more precisely, its main branches: mechanics, thermodynamics, hydrodynamics, aerodynamics, dynamics, and others. Therefore, there is no doubt that meteorology may rightfully be called physics of the atmosphere. According to the International Union of Geodesy and Geophysics (IUGG), meteorological science studies only a layer of the atmosphere in a shallow vertical space up to a height of 20 km. The rest of the atmospheric studies belong to so-called "*Aeronomy*". Aeronomy is the science of the upper part of the atmosphere, where dissociation and ionization are very important (Brasseur and Solomon 2005). This kind of division is obviously not appropriate because the processes in the atmosphere are interconnected. Thus, it is impossible to observe separately the wave motions that characterize the dynamics of the mesosphere and the lower thermosphere from the wave motions in the lower atmosphere, because these are interrelated motions. Or, atmospheric discharges in the troposphere result in spectacular electrical phenomena, such as blue jet, elves, and ring discharges (see Fig. 2.2).

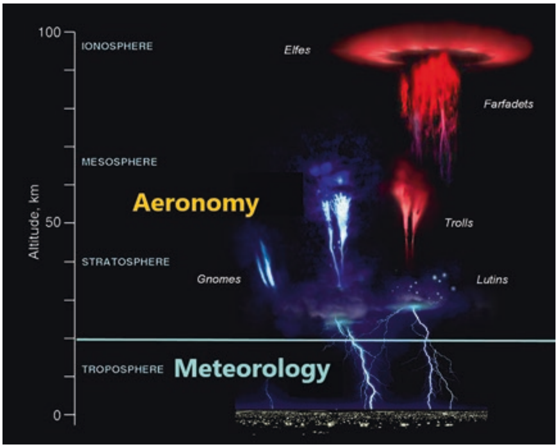

Fig. 2.2 The electrical discharges in the troposphere are related to that in stratosphere and mesosphere

2.4 Classification of Meteorology

Dynamic meteorology, physical meteorology, and applied represent three scientific branches focused on studying the fundamental principles, the atmospheric processes, and the application in various sectors.

Dynamic Meteorology. Dynamic meteorology studies the movements in the atmosphere using the laws of the dynamics and thermodynamics of the atmosphere. These basic laws applied to the atmosphere become very complicated, especially in situations where clouds are formed. The main task of this branch of meteorology is to provide a theoretical basis for understanding the effects of atmospheric movements on the weather and climate from the smallest to the largest. Within dynamic meteorology, a numerical weather forecast is being developed which solves the equations of the dynamics and thermodynamics of the atmosphere by numerical methods, since in most cases their solutions cannot be found by analytical methods. It is the science of dynamics of the atmospheric processes (Fig. 2.3). It is a fundamental discipline in meteorology where the complex meteorological processes and phenomena are interpreted in theoretical way using the fundamental principles of fluid dynamics and thermodynamics. The physical quantities that characterize the state of the atmosphere are temperature, density, pressure, etc.

Physical meteorology explains the physical processes in the atmosphere, such as solar radiation, absorption, reflection and scattering, and outgoing terrestrial radiation. It studies the cloud physics (see Fig. 2.4), aerosols, precipitation, atmospheric moist processes, and near surface processes such as mixing, turbulence, and friction.

Applied meteorology includes a wide range of individual scientific disciplines which deal with a specific application of meteorology in many sectors: transport, health, environment, agriculture, and water economy (Fig. 2.5). In the following

Fig. 2.3 Dynamical meteorology

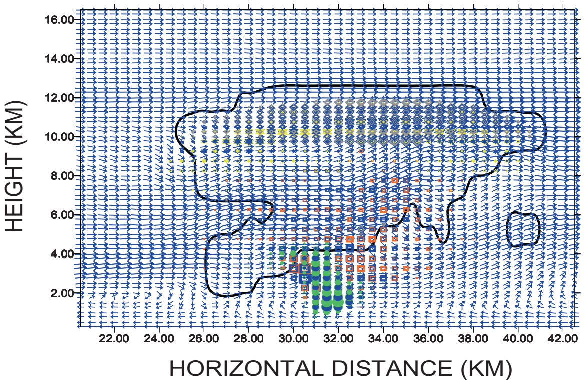

Fig. 2.4 Physical meteorology

text, a brief description of each individual branch of the applied meteorology is given.

Synoptic meteorology is a branch of applied meteorology whose name is derived from the Greek word synopsis, which is understood. This branch of meteorology aims to forecast the weather. Synoptic meteorology analyses processes in the atmosphere based on measurements and observations recorded on geographic maps for places where measurements are made at a point in time, so-called synoptic maps. The modern weather forecast is increasingly based on a numerical forecast, but still with the decisive role of man in interpreting the synoptic situation that arises based on drawn diagnostic maps and diagrams, products of different numerical models, and radar and satellite images. *Synoptic meteorology* is the science of weather prediction (e.g. Lackmann 2012; Barry and Carleton 2001; Bott 2012). The basic method is synoptic, and it consists of an analysis of atmospheric phenomena and processes with the help of special maps with a scale on which the data are caused by meteorological observations carried out at different points on Earth. These maps are called synoptically charts. The name derives from the Greek word “synopsis” which means “entry”, thus enabling to obtain a representation of time over a wider region, drawn on the map, and to provide forecasts for future situations.

Fig. 2.5 Applied meteorology

Climatology refers to a science that is focused on formation of climate in different geographical regions. "Climate" refers to a long-term specific regime of meteorological weather of a given area designated by the influx of solar radiation, the properties of the Earth's surface, and the atmospheric circulation.

Agriculture meteorology is a discipline which has studied the impact of meteorological factors on the state of agriculture.

Biometeorology or *medical meteorology* is a discipline that studies the impact of weather on human health. It is a separate branch of applied meteorology and studies the human sensitivity to various atmospheric conditions. It is an interdisciplinary activity in science that studies the interactions between the biosphere and the atmosphere. Human biometeorology is a part of biometeorology, which deals with the study of weather and climate impacts on human health. The title biometeorology is sometimes replaced with the term bioclimatology that studies the climate impact on health.

Radar and satellite meteorology is a discipline that deals with teaching methods, techniques, and tools for radar observation and identification of clouds and cloudy systems and methods and devices for satellite monitoring of the atmosphere and the processes (Raghvan 2003).

Hydrometeorology is a branch of applied meteorology, which refers to the hydrological cycle, water balance, and statistics of intensive rainfall. Hydrometeorologists prepare and issue forecasts of accumulated (quantitative) precipitation, the rainfall and snowfall intensity, and emphasize the areas with potential flash flooding.

Aeronautical meteorology is a branch of applied meteorology that refers to the impact of weather on air traffic management. It is important for pilots to understand the implications of weather on their flight planning and on their aircraft, as outlined in the manual for air information.

Environmental meteorology is the study of the physics, dynamics, and chemistry of the interactions of the Earth's atmosphere and the urban built environment and the provision of meteorological services to the populations and institutions of metropolitan areas. Environmental meteorology mainly analyses industrial pollution dispersion physically and chemically based on meteorological parameters such as temperature, humidity, wind, and various weather conditions important for architecture including analysis of how the construction phase of industrial and/or commercial buildings, residential housing, highways, bridges, and towers can be best accomplished during sustained periods of good weather.

Technical meteorology deals with the application of meteorology in separate technical branches of the economy, electric power economy, construction, transport, tourism, and other disciplines.

Weather modification is a branch of applied meteorology dedicated to human activity to affect the atmospheric processes which characterize the weather. There are standard methods and modern techniques to artificially influence and modify weather, also used for rain augmentation, hail suppression, fog dispersal, control of electrical discharges, and icing prevention. There is also another applied discipline as forensic meteorology.

Forensic meteorology is meteorology, focused on scientific studying of weather, applied to the process of reconstructing weather events for a certain time and location. This is done by acquiring and analysing local weather reports such as surface observations, radar and satellite images, and other data.

2.4.1 Classification Based on the Studied Area

Depending on the region or area where certain atmospheric processes and phenomena are manifested, meteorology is divided into tropical meteorology, polar meteorology, marine or ocean meteorology (oceanography), and mountain meteorology (Fig. 2.6).

Tropical meteorology studies the structure of the atmosphere and its behaviour in areas around the equator, roughly between 30 deg. north and south latitude. Weather and climate of tropics include phenomena such as northeast and southeast trade winds, hurricanes, Intertropical Convergence Zone, jet streams, monsoon, and El Niño-Southern Oscillation. Over the tropics more energy from the sun arrives than it is lost in space (infrared radiation). The opposite occurs in high latitudes from 30 deg. lat to the poles. The excess energy from the tropics is carried by the winds in high latitudes, by means of vertical circulation known as *Hadley cell.*

Polar meteorology is a branch of meteorology which explores the weather and climate in high latitudes of the Earth. In polar regions, the Sun never rises far above the horizon, and it remains constant throughout the year, so that snow and ice can remain for long periods even at lower positions. Meteorological processes that occur have exceptionally local characteristics in large proportions of the two Polar Regions.

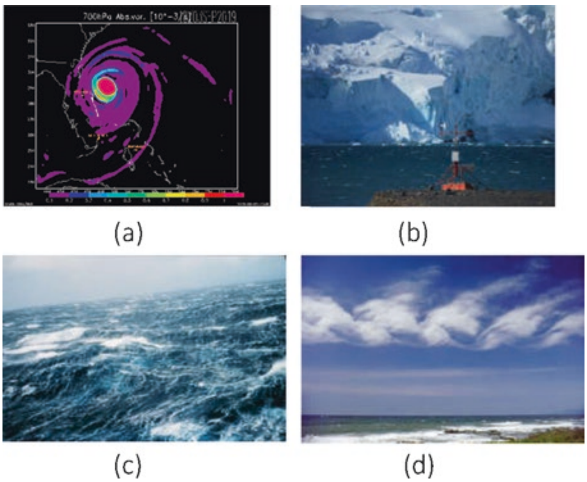

Fig. 2.6 Region based classification.
(**a**) WRF-ARW forecast of hurricane Dorian 3 Sep 2019; (**b**) Landscape-Weather Station, Antarctica Research; Credit: Pixabay; (**c**) North Pacific storm waves as seen from the NOAA M/V Noble Star, Winter 1989; Credit: NOAA; (**d**) A shot of the unusual Kelvin-Helmholtz waves in cirrus cloud; Credit: Astronautilus; https://creativecommons.org/licenses/by/2.0/deed.fr

Marine or ocean meteorology mainly deals with the study of oceanic areas, including islands and coastal regions. This discipline is a practical application for the Earth's needs and air navigation through the ocean.

Coastal meteorology studies meteorological phenomena in the inner part of the mainland, about 100 km from the coast. In meteorology, the determination of the processes in a coastal zone is based on detailed knowledge of maritime and land boundary layers and the interactions of the system atmosphere-sea. But, in order to understand the processes better, the atmospheric dynamics in large-scale circulation and coastal ocean should also be considered. Coastal meteorology is used for weather forecasting in coastal areas. This branch of meteorology helps to understand the physical, chemical, and biological aspects of ocean coastline.

Mountain meteorology is focused on the effects of mountains on the atmosphere, which are placed in all scale processes of movement. Alpine meteorology is a special branch of this scientific discipline that deals with the impact of the mountain system of central and southern Europe, approximately 805 km in length and 161 km in width, which stretches from the Mediterranean Sea through northern Italy and southeastern France, Switzerland, southern Germany, and Austria and in the northwestern part of the Balkan Peninsula. Based on the studied layer in the atmosphere, meteorology is divided into planetary boundary layer meteorology and aerology.

The boundary layer meteorology. It is the branch of meteorology that studies the processes in the layer of air which is directly above Earth's surface and is known as the atmospheric boundary layer (ABL) or planetary boundary layer (PBL) (Fig. 2.7). The effects of surface heating, cooling, and friction cause turbulent mixing within the layer of air. Considerable heat flux, mass, or momentum in a time scale less than a day is transported with a turbulent movement. Boundary layer meteorology includes the study of all types of surface boundaries: Earth-surface environment, including oceans, seas, and lakes and urban and non-urban lands.

Fig. 2.7 Boundary layer meteorology

2.4.2 Classification According to the Scale of Processes

The atmosphere has a wave nature, characterized by different spatial and temporal distribution of phenomena and processes (Lin 2007). Therefore, when studying the atmosphere, meteorology can be divided into different areas of significance and importance, depending on the spatial and temporal field of interest. One extremely important branch of this scale is climatology. The time scale of meteorology, hours to days, is divided into micro-scale, meso-scale, synoptic-scale, and global-scale meteorology (see Fig. 2.8).

Micro-scale meteorology studies a small-scale atmospheric process, associated with the interaction of the atmosphere and Earth's surface. In these processes surface transport and exchange of energy, heat and humidity at the ground layer of atmosphere, convection, microphysical processes in clouds, turbulent processes in

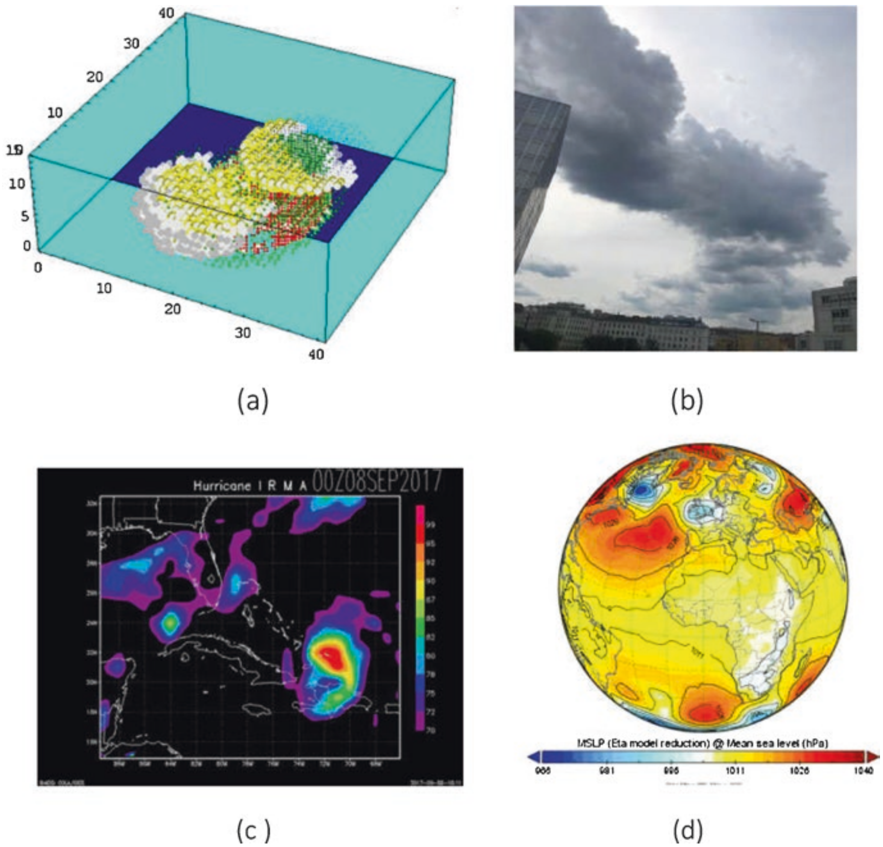

Fig. 2.8 Classification based on the scale of the processes. (**a**) Numerical simulation of cloud-aerosol chemistry; (**b**) rotor clouds as the mesoscale instability phenomena; (**c**) numerical simulation of hurricane Irma; (**d**) global mean sea level pressure in hPa

the atmosphere, and the processes of diffusion are included. Spatial scale of these processes is with a horizontal dimension less than 2–10 km, with time scales between 15 min and 1 h. *Meso-scale meteorology* is a branch covering atmospheric processes which have horizontal scales ranging from micro-scale to synoptic scale from 20 to 200 km approximately, with a vertical scale that starts at the Earth's surface and covers the atmospheric boundary layer, the troposphere, tropopause, and the lower part of the stratosphere. Time scale of meso-scale processes may be from a few hours to less than 1 day, until the end of the life cycle of the phenomenon. The meso-scale phenomena of special interest are storms, MCS, squall lines, convective bands, frontal precipitation in tropical cyclones and extratropical cyclones, and topographic established weather systems, such as mountain waves and winds from the sea and land.

Synoptic-scale meteorology is basically a more dynamic field that refers to the horizontal coordinates in terms of time. It includes phenomena such as cyclones, extratropical cyclones, baroclinic troughs and ridges, frontal zones, and, to some extent, jet currents. All these phenomena are usually drawn on weather maps, to define specific time. Minimum horizontal scale of synoptic phenomena is limited in space between the surface and observing stations.

Global-scale meteorology or planetary scale meteorology encompasses the global atmospheric circulation, long waves, baroclinic waves, synoptic cyclones, radiation, and surface processes of heat exchange and moisture between the oceans and land. The study of temporal forms in this area covers the transport of heat from tropics polewards. Also, the oscillations in a large scale are of extreme importance. These oscillations have periods that are greater than the annual seasonal cycle. Global scale directs the thresholds of perception of the meteorology towards climatology. The traditional definition of climate is turned down in a larger time scale, explaining how the global oscillations caused climate and weather disturbances in synoptic and mesoscale.

2.5 The Modern Term of Meteorology

Today we are living in an era of climate change and global warming. We are witnessing an increased frequency of atmospheric natural disasters that endanger our planet Earth. Development and application of new technologies contribute to obtaining better quality knowledge and information on the status and behaviour of the atmosphere of our planet and the processes occurring in it. Modern view of the meteorology (Helmis and Nastos 2012) almost transcends the conventional definition of meteorology as the *"Physics of the Atmosphere"*. Meteorology, as a subdiscipline of geophysics, grows into an interdisciplinary science, which is widely devoted to the study of energy and planetary processes and interactions that take place between the atmosphere, hydrosphere, lithosphere, and biosphere of planet Earth and greatly affects the global climate system.

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Chapter 3
Historical Background

3.1 Aristotle's Meteorologica

Occurrences in the atmosphere have long been a subject of human interest. As with other sciences, it is almost impossible to precisely determine the beginning of the development of meteorology (e.g. Ćurić 2006; Neves et al. 2017; Lutgens and Tarbuck 2009; Spiridonov and Ćurić 2010). One should make a distinction between meteorology as a natural science and meteorology as a separate "branch of knowledge". It seems that the term "meteorology" originates from 340 BC when the Greek philosopher Aristotle (Fig. 3.1) wrote a book on human philosophy titled "Μετεωρολογικά" (Greek) "*Meteorologica*" or "*Meteora*" (Latin). This book represented the philosophical study of the atmosphere at that time, including different meteorological elements and weather phenomena (e.g. humidity, clouds, rain, snow, wind, hail, lightning, and thunder). Also, other areas such as astronomy, geography, and chemistry were also included. The book was titled "*Meteorologica*" because each particle that was falling from the sky or that was suspended in the atmosphere was called "meteor". Today, we make a distinction between "meteors" (extra Earth meteors) and hydrometeors (particles of water or ice in the atmosphere). The Greek word "meteors" refers to something "high in the sky", which is located between Earth and the universe, while the word "logos" means study.

Although there were no instruments yet, the ancient Greeks collected data on the state of the atmosphere, primarily wind and precipitation, and various atmospheric phenomena were described and interpreted. Greek philosophers seemed to understand the hydrological cycle, while water vapour, since it was invisible, was less understood. Aristotle's disciple Theophrastus wrote "Weather Signs" containing a set of prognostic rules derived from celestial phenomena. In *"Meteorologica"* Aristotle tried to explain weather phenomena in a philosophical way. Although most of his claims were wrong until the seventeenth century, many Aristotle's original ideas were not scientifically recalled.

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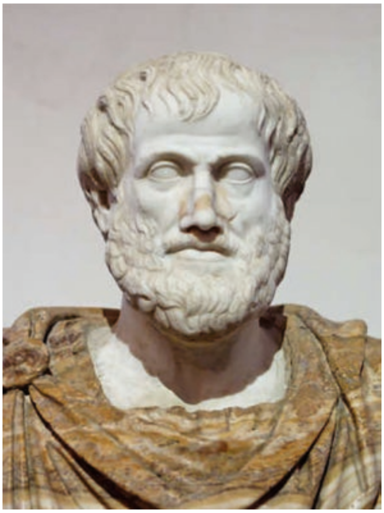

Fig. 3.1 Bust of Aristotle (384–322 BC)

3.2 Early Research Period

The origin of meteorology dates since its beginning in 3000 BC (Teague and Gallicchio 2017), through development as a legitimate, natural science dating from late sixteenth century. Meteorology has been recognized as a natural science only since the end of the 16th century. At that time, it was clear that the speculations of natural philosophers were inadequate and that greater knowledge was necessary for further understanding of the atmosphere. Certain instruments that would likely measure properties of the atmosphere (e.g. temperature, humidity, pressure) were necessary, but such instruments were never found until the end of that period. The first of such discoveries was the hygrometer. Greek philosophers seemed to understand the physical basis of the circulation of water through land-atmosphere system, even in the absence of the instruments. Because the water in the gas (vapour) state is invisible, these processes were less understandable. Questions pertaining to the properties of this "invisible water vapour" led to the finding of an instrument for measuring humidity by the German mathematician Cardinal Nicolas de short, around 1450. The second invention of that kind was the water thermometer. The discovery was attributed to Galileo Galilei (Fig. 3.2) in 1593, although the exact date of his finding is not certain to this day. The ideas and the inventions of Galileo motivated other scientists at that time to continue the search and the discovery of other instruments for measuring atmospheric phenomena. Another invention in the seventeenth century was instrument for measuring atmospheric pressure, called "barometer". The "barometer" is known as an instrument of Torricelli Evangelista, an Italian mathematician who studied in the class of Galilei.

In 1643, Torricelli and his student Vincenzo Viviane constructed a vacuum tube and used mercury for measuring the weight of the air column (Fig. 3.3). The first anemometer for measuring wind speed was constructed in 1667 by Robert Hock,

Fig. 3.2 Galileo Galilei (1564–1642)

Fig. 3.3 Torricelli Evangelista (1608–1647)

while Horace de Saucer completed this list of important meteorological instruments in January 1780, with the construction of fibrous hygrometer as an instrument for measuring the humidity of the air. In 1765, daily measurements of air pressure were conducted; the moisture content of the direction and speed of wind were determined. For the first time, this is done by the French scientist, Lavoisier Laurent, who stated: "With all this information it is almost always possible to provide the weather forecast with reasonable accuracy, one to two days in advance". However, things were not as simple as Lavoisier thought. Since 1872 the Government British Meteorological Office had prepared daily weather maps collecting the data and drafting on a simple weather chart. Later, a new graphical tool was developed to enable more rapid transmission of information. In that period, isobars-drawn lines on a weather map, connecting points of equal barometric pressure and isotherms-connecting lines having equal temperature were found. Other graphic devices were also developed for characters for the display of wind direction and strength, as well as lines that describe the intersections of warm and cold air masses. In 1854 a French warship and 38 merchant ships sank in fierce storm near Krim's port of Balaklava. The Director of the Paris Observatory was asked to investigate the temporal disaster. With the analysis and verification of meteorological reports, it was noticed that the storm occurred 2 days before the ships sank and passed through Europe from the southeast. If there was a system for monitoring the weather at that time, ships would be notified of a storm surge. As a result of these findings, a national storm warning service was established in France. This period is recognized as the beginning of the modern meteorology. In 1862 Lord Kelvin, Scottish mathematician and physicist, suggested an absolute temperature scale, as the important finding, which allowed further application in research and study of temperature characteristics of the atmosphere. Instruments continue to develop during the eighteenth and nineteenth centuries, with the full support of new technological achievements that impact on the deepening of our knowledge of the atmosphere. The invention of telegraph in 1843 allowed routine transmission of the weather observations and the observations installed across the globe. Using these data, weather maps and tracks from the surface of the wind were roughly drawn, and with their help storm systems could be identified and spatial scales could be studied for a longer time. Among many other advances, mainly related to the progress in physical science was the discovery of Robert Boil, the dependence of the volume of the gas pressure, which led to the thermodynamics and experiments with lightning by Benjamin Franklin. The first correct explanation essential to the global atmospheric circulation, whether based on a study of trade winds of George Hadley (Fig. 3.4) written in 1735, which contributed to the mean tropical cell zonal atmospheric circulation is referred as "Hadley cell". In 1835, physicist Gaspard-Gustave de Coriolis

Fig. 3.4 George Hadley (1685–1768)

found that the rotation of Earth causes power to depend on the velocity of bodies in a reference composition of non-rotating Earth. Synoptic observations at the time showed that there were some difficulties for the establishment of certain characteristics associated with weather, such as clouds or winds. These difficulties were overcome when Luke Howard and Francis Beaufort introduced systems for cloud classification (1803) and wind speeds (1806), respectively. A real milestone in that time was the invention of the telegraph in 1843 which has provided exchange of information within invaluable time speed.

3.3 Modern Research Period

At the beginning of the twentieth century, theoretical studies of atmospheric phenomena were usually performed analytically, taking the basic equations of fluid dynamics which govern atmospheric processes, simplifying them to the neglect of more terms, and seeking solutions to these equations (Thaxter 1990). Vilhelm Friman Koren Bjerknes (Fig. 3.5) is a pioneer in the modern practice of weather forecasting. He was especially famous in his studies of hydrodynamics and thermodynamics and their association with the atmospheric motion. For example, Bjerknes developed the model that explains the formation, intensification, and dissipation, i.e. the life cycle of cyclones at high latitudes, by introducing the idea of frontal systems as closely defined boundaries between two air masses (Bjerknes 1921).

In 1922, an English mathematician, physicist, meteorologist, psychologist Lewis Fry Richardson (Fig. 3.6) made a pioneering work in developing the first numerical weather prediction experiment. He is also noted for his pioneering work developing a method for solving a system of linear equations known as modified Richardson

Fig. 3.5 Vilhelm Bjerknes (1862–1951). (Credit: Bjerknes Family [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

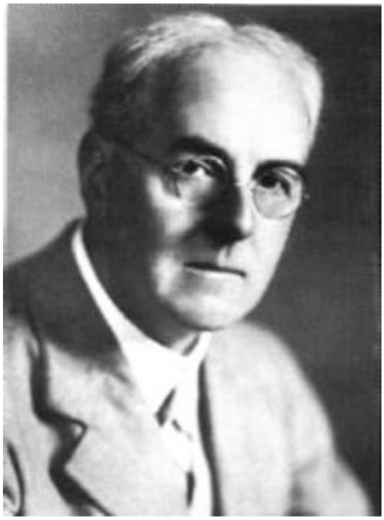

Fig. 3.6 Fry Lewis Richardson (1881–1953)

iteration. Richardson was several decades ahead of his time in what he attempted to do. At the time of the World War I, computational weather forecasting was impractical for several reasons (e.g. lack of available observations, instabilities of the algorithms for solving the atmospheric equations, inadequate knowledge of the balance of the atmosphere, and massive volume of computation required to advance the numerical solution). Later, John von Neumann, Jule Charney, and Ragnar Fjørtoft made the first numerical weather prediction by computation of a solution of a simple equation, the barotropic vorticity equation (BVE), on the only computer available, the ENIAC. Starting in the 1950s of the nineteenth century, the numerical experiments with computers became feasible and practical (Lindzen et al. 1990). The first weather forecasts derived this way used bar tropic (single-vertical-level) models, which could successfully predict the movement of Rossby waves in large scale, a general model of atmospheric cyclones and anticyclones.

The theory of chaos, i.e. nonlinearity of atmospheric processes, has great significance for the development of meteorology, especially in the description of the limitations of the forecast period; that fact is an integral part of atmospheric modelling. The chaotic nature of the atmosphere was first studied by Edward Lorenz.

In 1960, Lorenz reveals the *theory of chaos* and the well-known "*butterfly effect*". A plot of Lorenz's strange is shown in Fig. 3.7. In his theory, Lorenz is linking the 1-min-long swings of the wings of a butterfly in a place with great disturbances which later would be caused elsewhere. Faster technological development in the 1960s allows launching of the world's first weather satellite (TIROS-1), which was equipped with a camera. Today, weather satellites pass from pole to pole. Geostationary satellites are positioned in a fixed position above Earth and constantly monitor a part of the globe.

The development of advanced remote sensing techniques such as radars, satellites, lidars, sodars, and other technology continues so that meteorological satellites

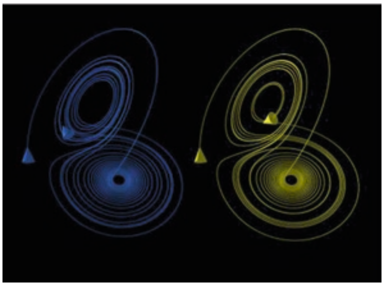

Fig. 3.7 Two Lorentz orbits

of the second generation (MSG-2) are designed to monitor the advantage of the new technology and to improve the already successful and proven design of the original (METEOSAT) satellites. The development of modern satellite technology creates tools that are applied in the monitoring the weather and climate, the atmospheric phenomena, and processes that are manifested at different spatial and temporal scales and which substantially affect our daily lives and our daily activities. Today, automated weather stations (AWS) send balloons into the atmosphere that contain so-called radiosondes. These instruments can measure atmospheric conditions using the radio to return information back to the station. Of course, weather stations now also use radars. New scientific and technological achievements in the twenty-first century allowed rapid progress in the development and application of computer models for weather forecasting, water, climate, and environment. More detailed description of the historical development of meteorology, early research period until the modern period of development and application of meteorological science, can be found in one of the few monographs which are a historical treasure for development of meteorological practice and science.

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Chapter 4
Structure and Composition of the Atmosphere

4.1 Earth Spheres

Earth is a complex system, which covers four main spheres given below and shown in Fig. 4.1:

- *Atmosphere* (gas layer)
- *Lithosphere* (solid Earth part)
- *Hydrosphere* (aqueous part)
- *Biosphere* (living world)
- *Cryosphere* (ice part)

Spheres are interrelated and constitute an integral system, which conducts a very important process. The exchange of energy that constantly occurs between the Earth's surface and the atmosphere, and between the atmosphere and the surrounding space, creates effects that define the nature of weather. The Earth system covers complex and continuous interaction between lithosphere, hydrosphere, atmosphere, and biosphere. The two main energy sources that operate this system are:

1. Solar energy, which is forcing the external processes that occur on or above Earth's surface
2. Thermal energy, which comes from the Earth's interior, i.e. heat to be supplied at the formation of the planet and heat that is released in radioactive decay.

4.2 Basic Characteristics of the Atmosphere

The atmosphere represents a thin layer of gases, around the Earth's surface which is retained by Earth's gravity. Like the atmospheres of other planets, the Earth's atmosphere plays a key role in the exchange of energy between the Sun and the surface

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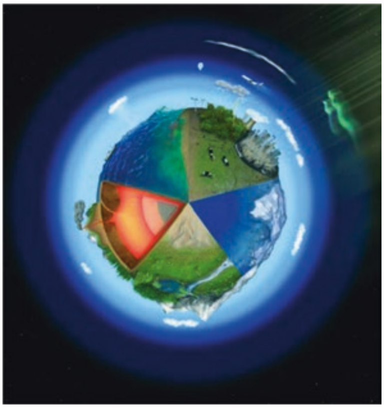

Fig. 4.1 Earth as a complex system. (Credit: ESA/AOES Medialab [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)])

of the planet itself, as well as between different areas of the planet. This energy exchange maintains a thermal balance and determines the climate of the planet. The Earth's atmosphere, of course, is unique in that it is also directly supported by the huge ocean surface, which, through appropriate processes, places this system in a favourable environment for the living world. The atmosphere is a fluid medium in which a variety of movements occur, from turbulent (in the area of about a few meters) to circulations that have dimensions as far as the planet itself (Lutgens and Tarbuck 2009). These air movements influence other components (water vapour, ozone, clouds) that significantly change the global energy balance of the atmosphere.

4.3 Origin of the Atmosphere

The origin of the atmosphere is still subject to different interpretations (Mölders and Kramm 2014). Many of the scientists, based on their knowledge of gases in the universe, believe that Earth's first atmosphere was composed of the following gases: helium, hydrogen, ammonia, and methane. Others believe that the first atmosphere was probably composed of gases released from volcanoes emitting warm inner core. The following are included in this group of gases: nitrogen ($N_{2),}$ carbon dioxide ($CO_{2),}$ hydrogen ($H_{2),}$ carbon monoxide (CO), and water vapour (H_20). It is anticipated that the second Earth's atmosphere is generated as a result of heating and differentiation. It probably consisted of the same gases that are released today by volcanoes: carbon dioxide, nitrogen, water vapour, hydrogen, and other gases that are present in the form of tracers. Planetary differentiation caused the lighter elements to rise to the other top layers of the Earth and thus caused the outpouring of the lighter gases from the interior of the planet. It is assumed that the lighter gases form the atmosphere and the oceans.

4.4 Chemical Composition of the Atmosphere

The atmosphere is a mixture of solid, liquid, and gaseous constituents. Gases in the atmosphere are classified as constant (the concentration remains constant) or variable where the concentration changes with time (Schumann 2012). Permanent gases include oxygen, nitrogen, neon, argon, helium, and hydrogen. Mostly present of the permanent gases are nitrogen (78.1%) and oxygen (20.9%) and argon (0.934%). Rest of atmospheric gases are known as tracers or trace gases, because they are found in very small concentrations or benchmark (Fig. 4.2). Trace gases include carbon dioxide, ozone, and specific gases such as argon, neon, helium, krypton, and xenon, which are very inert and essentially involved in any chemical transformation within the atmosphere. The atmosphere also contains sulphur, fluorocarbons, and dust and ice particles.

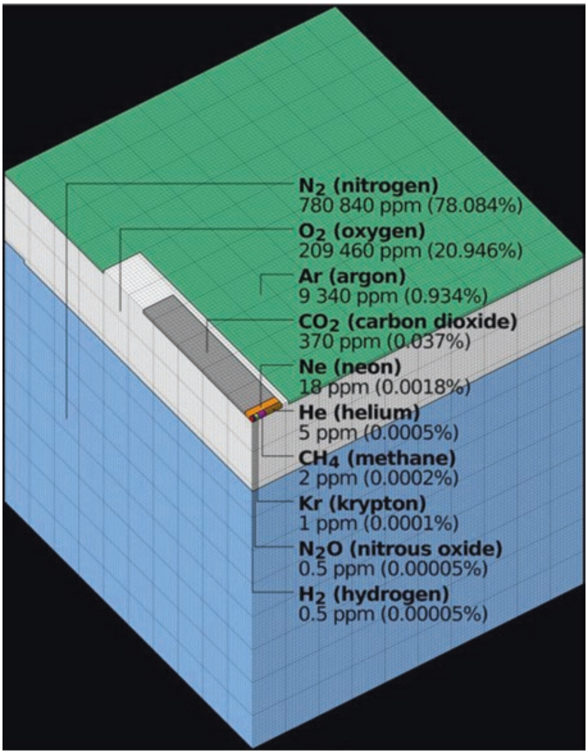

Fig. 4.2 Chemical composition of the atmosphere. (Credit: Cmglee [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)])

4.5 Significant Atmospheric Gases

4.5.1 General Facts About the Atmosphere

The atmosphere consists of a dynamic mixture of gases surrounding the Earth. The two main gases, nitrogen and oxygen, make up the largest volume part of the atmosphere. They are important for the maintenance of life and functioning of many processes near the Earth's surface. In the atmosphere there also exist "secondary gases" which play equally important role in the Earth's system (Jacob 1999). The gases which have a significant impact on heat budget and the availability of moisture on Earth are called secondary gases.

The atmosphere has layer structure due to the presence of characteristic thermal layers, which appear as a result of changes in air temperature with height. The following chapters will examine the importance of primary and secondary emissions and describe the vertical thermal structure of Earth's atmosphere. Atmospheric gases are divided into permanent and variable gases. Permanent gases are nitrogen, oxygen, argon, neon, and helium (Table 4.1). Variable gases include water vapour, carbon dioxide, methane, nitric oxide, ozone, and matter chlorofluorocarbons (Table 4.2).

Nitrogen: Nitrogen is the most abundant gas in the atmosphere. Large amounts of nitrogen enter the atmosphere through volcanic eruptions, biomass burning, and nitrification. Nitrogen is released from the atmosphere and deposits of the Earth's surface with nitrogen-retainer bacteria and lightning. Nitrogen serves as an important nutrient for plants.

Table 4.1 Permanent gases in the atmosphere

Gas	Symbol	Percent of air (by volume)
Nitrogen	N_2	78.08
Oxygen	O_2	20.95
Argon	Ar	0.93
Neon	Ne	0.0018
Helium	He	0.0005

Table 4.2 Variable gas emissions

Gas	Symbol	Percent of air (by volume)
Water	H_2O	0–4%
Carbon dioxide	CO_2	0.0380%
Methane	CH_4	0.00017%
Nitrous oxide	N_2O	0.00003%
Ozone	O_3	0.000004%

Oxygen: Oxygen is the second dominant gas in the atmosphere. It is a two-atomic molecule, which is released into the atmosphere as unproductive products in the process of photosynthesis. During photosynthesis carbon dioxide and water react with the help of sunlight, which form oxygen and glucose. Oxygen is consumed during the process of chlorophyll assimilation (respiration). Oxygen chemically combines with glucose to obtain ATP together with carbon dioxide and water as waste products.

Water vapour: Water vapour represents the most important component of changing emissions. It is a natural greenhouse gas. Greenhouse gas prevents the infrared radiation coming from the Sun. Embedded infrared radiation heats the Earth. This phenomenon is known as the greenhouse effect. Water vapour, absorbs heat radiated from the earth's surface in the lower atmosphere. In this way, water vapour plays a significant role in the redistribution of energy in the atmosphere. Water vapour is released from the atmosphere during precipitation and back into the atmosphere through evapotranspiration. Water vapour concentration in the atmosphere changes depending on the place and the time of the year. Larger values of water vapour concentration are evidenced over the oceans, seas, and tropical wet forests.

Carbon dioxide: Carbon dioxide is the fifth most abundant gas in the *atmosphere.* Carbon dioxide enters the atmosphere during volcanic eruptions, burning of biomass and fossil fuels, cement, and cutting wood. Carbon dioxide from the atmosphere is released during process of photosynthesis and absorption in the oceans. In the oceans, part of the carbon dioxide reacts with water to form carbonic acid. The concentration of carbon dioxide increased by about 25% in the last 300 years.

Ozone (O_3): Ozone is a natural component of the atmosphere that consists of three (3) oxygen atoms (Fig. 4.3). Ozone is an unstable molecule that quickly decays into oxygen molecule and another atom of oxygen. Ozone in the stratosphere protects the atmosphere from harmful ultraviolet radiation, known hazards to plants and animals, including humans. Ozone destruction causes harmful effects on people which include the risks of skin cancer, cataract, and adverse impacts on the immune system. At lower heights, ozone has different effects. Ozone can cause respiratory problems (smog contains significant gas ozone), and it is corrosive and reactive with building materials. Recently, ozone has been appointed as greenhouse gas. The percentage representation of nitrogen in unit volume is 78.08%. Oxygen is represented by 20.95% and argon 0.93%, while the smallest quantity contains neon and helium.

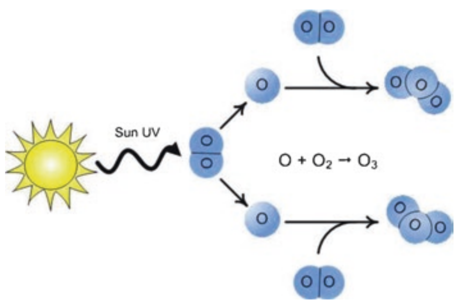

Fig. 4.3 Ozone formation

Water vapour is the most important variable gas with volume presence in the air that ranges from 0 to 4%. Percentage volume of carbon dioxide (greenhouse gas) is 0038% or 380 ppmv.

The remaining gases from Table 4.1 have much less volume presence in the atmosphere.

The present-day carbon dioxide (CO_2) concentration is about 365 parts per million volume (ppmv) or 365x10^{-6} mol/mol (Jacob 1999). Stratospheric values of CO_2 at heights between 11 and 50 km are around 5–12 ppmv. The concentration of variable gases changes over time. The gas field that surrounds the planet is divided into several concentric layers. Some (99%) of the total weight of the atmosphere is concentrated in the first 32 km above the Earth's surface.

4.6 Atmospheric Structure

The Earth's atmosphere is divided vertically into layers (e.g. Andrews 2010; Barry and Chorley 2010; Borngraber 2018; Mcdougal 2004; Saha 2008). Each has specific features characterizing the atmospheric status and its future behaviour. According to the vertical profile of temperature in the atmosphere, the following atmospheric layers are distinguished: troposphere, stratosphere, mesosphere, thermosphere, and exosphere (Fig. 4.4).

Troposphere. Troposphere is the layer of atmosphere closest to the planet characterized by the largest percentage (80%) of weight of the overall atmosphere. Temperature and water vapour content in the troposphere decrease rapidly with height. Water vapour plays a major role in regulating air temperature because it

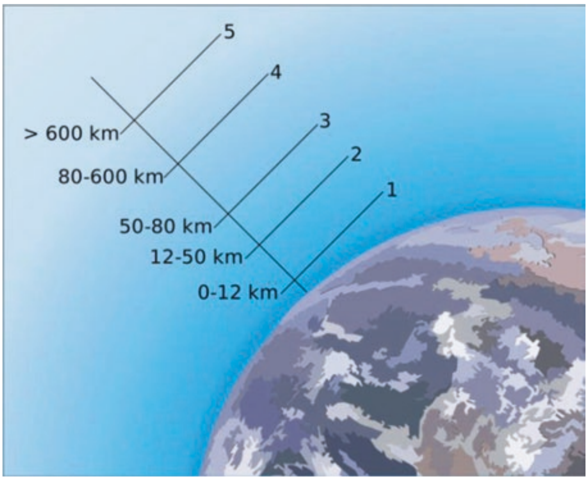

Fig. 4.4 Structure of the atmosphere. (This SVG image was created by Medium69.Cette image SVG a été créée par Medium69.Please credit this: William Crochot [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

absorbs the incoming energy from the Sun and thermal radiation from the Earth's surface. The troposphere contains 99% of water vapour in the atmosphere. Concentrations of water vapour change with geographical latitude.

They are greatest above the tropics, where they may be higher than 3%, and they reduce in the Polar Regions. All weather phenomena occur within the troposphere, although disturbances can be extended to these influences of the lower stratosphere. The term "troposphere" means "region of mixing". Inside the layers of this region, there exist strong convective currents of air. The ceiling layer, known as *tropopause*, is bordering from 8 km near the poles to 18 km above the equator. The height of tropopause also changes with the seasons, highest in summer and lowest during the winter.

Stratosphere. Stratosphere is the second major layer of air in the atmosphere. Stretching over the tropopause to a height of about (50 km) from earth's surface. Since the air temperature in the stratosphere increases with height, the stratosphere is stable, and there are no conditions for convection. Above the stratosphere, there is a thin layer called stratopause, in which the temperature remains relatively constant with height. Ozone plays a key role in regulating the heat regime of the stratosphere. The content of water vapour inside the layer is very low. The temperature increases with the concentration of ozone. Solar energy is converted energy into kinetic when ozone molecules absorb the ultraviolet rays, which result in heating of the stratosphere. Ozone layer is located between the heights 15 and 25 km. Approximately 90% of ozone in the atmosphere is in the stratosphere. Volume concentration of ozone in this region is 10 ppmv or 10 parts per million by volume compared with the concentration of ozone in the troposphere, whose approximate value is 0.04 ppmv. The pure ozone absorbs ultraviolet sun radiation in the range of wavelengths 290–320 nm (UV-B) radiation, which is harmful to life. Increased levels of ultraviolet radiation, which come from the Earth's surface, also affect life of plants negatively, with harmful consequences to the natural environment. Large values of sunlight, ultraviolet radiation would result in many biological effects and infrequent occurrence of cancer.

Mesosphere. Mesosphere is the layer that extends approximately from 50 to 85 kilometres above Earth's surface, and again it is characterized by decreasing temperatures. The lowest temperatures of the Earth's atmosphere appear at the top of this layer, known as "mesopause". This especially happens in summer near the poles. The mesosphere is sometimes called "minority area" because the layer is probably the least studied of the other atmospheric layers. The stratosphere and the mesosphere are sometimes known as *"middle atmosphere"*.

Thermosphere lies above the mesosphere. The temperature in the thermosphere generally increases with height, reaching 600–2000 K depending on the activity of the Sun. This increase in temperature occurs as a result of absorption of intense solar radiation by the limited amount of remaining molecular oxygen. Molecules of gas, at this extreme height, are widely separated. Above this layer with a height of 100 km from Earth's surface, the chemical composition of air becomes strongly dependent on height. The atmosphere is enriched with lighter gases (atomic oxygen, helium, and hydrogen). Going higher than 160 km, the main component of the atmosphere, atomic oxygen, can be found. At very large heights, the residents begin to be stratified according to molecular weight, due to gravitational separation.

Exosphere. Exosphere is the most distant atmospheric layer of the Earth's surface. In the exosphere, the molecules can go to space in the ascending path (provided they are moving fast enough) or to divert back to Earth with the help of gravity, with little probability of collision with other molecules.

The height of the lower boundary, known as "thermopause" or "exobase", ranges from about 250 to 500 km depending on solar activity. The upper level boundary may theoretically be determined by the height (half distance to the Moon, approximately 205.680 km) and the impact of solar radiation pressure of the hydrogen atom speeds exceeding the Earth's force of gravity. Exosphere observed from space as Earth's corona is seen ranges about 102,840 km from the Earth's surface. Exosphere represents a transitional zone between Earth's atmosphere and interplanetary space.

4.6.1 Thermal Structure of the Atmosphere

The layers of the atmosphere are characterized by changes in temperature. These changes are primarily resulting from absorption of solar radiation, visible light at the surface, near ultraviolet radiation in the middle part of the atmosphere, ultraviolet radiation, and distance in the upper atmosphere. The vertical temperature distribution (stratification curve) and the dew point temperature (moisture curve) at several characteristic layers of the atmosphere are shown in Fig. 4.5.

4.7 Magneto-electronic Structure

Based on the behaviour and the number of free electrons and other charged particles, the upper atmosphere is divided into regions: ionosphere, magnetosphere, and plasmasphere.

Ionosphere. Ionosphere depends on the atmospheric effects of radiation expansion which occurs as a result of the presence and the change in the concentration of free electrons in the atmosphere.

The plasmasphere is not spherical but a doughnut-shaped region (a torus) with the hole aligned with Earth's magnetic axis. The Earth's plasmasphere is made of just that, a plasma, the fourth state of matter. This plasma is composed mostly of hydrogen ions (protons) and electrons. It has a very sharp edge called the plasmapause. The outer edge of this doughnut over the equator is usually some 4–6 Earth radii from the centre of the Earth or about 19,000–32,000 km above the surface. The plasmasphere is essentially an extension of the ionosphere. Inside of the plasmapause, geomagnetic field lines rotate with the Earth. The inner edge of the plasmasphere is taken as the altitude at which protons replace oxygen as the dominant

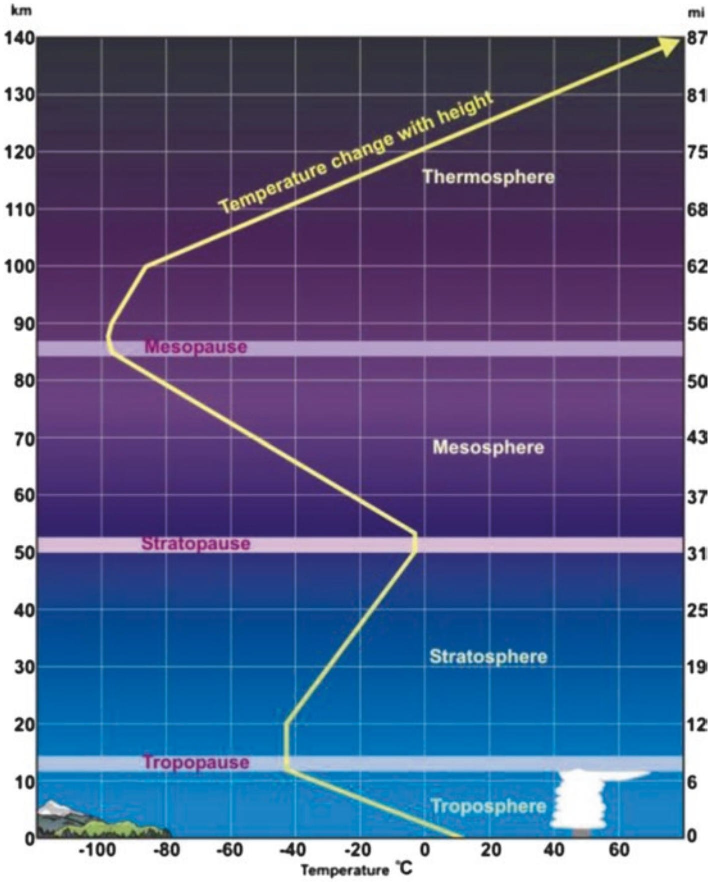

Fig. 4.5 Standardized temperature profile

species in the ionosphere plasma which usually occurs at about 600 miles (1000 km) in altitude. The plasmasphere can also be a structure within the magnetosphere.

Magnetosphere. Outside the plasmapause, magnetic field lines are unable to corotate because they are strongly influenced by electric fields of solar wind origin. Magnetosphere is a cavity in which Earth's magnetic field is limited by the solar wind and magnetic field interplanetary magnetic field (Fig. 4.6).

The outer boundary of the magnetosphere is called magnetopause. Actually, it is the boundary between the planet's magnetic field and the solar wind. The magnetopause is typically located at about 56,000 km above the Earth's surface.

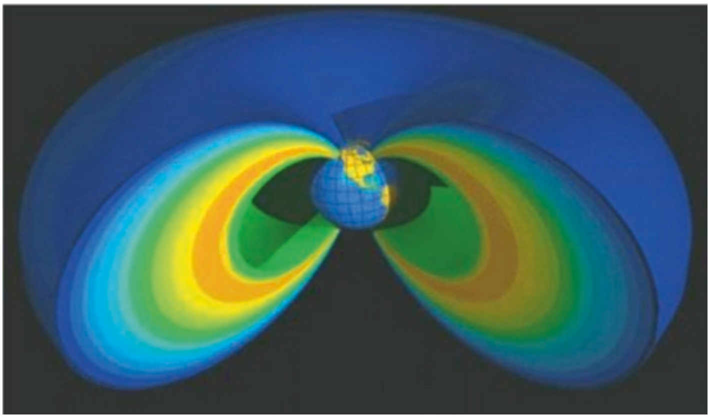

Fig. 4.6 Magnetosphere

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Chapter 5
Energy and Radiation

5.1 Basic Features

The energy emitted by the Sun provides almost (99.9%) of the total energy that heats the Earth's surface. It warms the Earth and the surrounding atmosphere, thereby supporting life. The main source of energy is solar energy that serves as a main driving force of the atmospheric circulation and the physical processes on planet Earth (Visconti 2016):

- Heating the atmosphere, water, and land of the Earth's surface.
- Major mechanisms for the general atmospheric and ocean circulation.
- Formation of air masses and atmospheric systems.
- Generating the winds and ocean currents.
- Photosynthesis that affects the overall life of our planet.
- Changes the atmosphere, hydrosphere, and lithosphere.
- It is the most important external factor in creating climate.

According to some estimates, the total amount of thermal energy is about 300 times lower than the quantity of solar energy that arrives daily in the upper limit of the atmosphere. Heat that comes from the Earth's interior to the surface of the Earth is about 5000 times lower than the heat that arrives from the Sun. The large changes in solar heat drive the winds in the atmosphere and currents in seas and oceans. Therefore, first, it is necessary to understand the way the Sun heats the Earth, and this warming is changing the geographical space and time. For this purpose, we primarily study the heat and the temperature as variables and the mechanisms for their transfer.

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https://doi.org/10.1007/978-3-030-52655-9_5

5.2 Radiation Laws

Black body radiation. Any substance (solid, liquid, or gas) emits radiation according to its absolute temperature, measured in units of Kelvin (K = °C + 273.15). The efficiency at which radiation is emitted varies with the substance. However, at any temperature, there is an upper limit to how much radiation can be emitted. A black body is a hypothetical body that:

1. Emits radiation at the maximum intensity possible for every wavelength
2. Completely absorbs all incident radiation (hence the term "black")

Planck's law. Describes the amount of radiation emitted by a black body at each wavelength as a function of temperature. Monochromatic irradiance is given with:

$$E_{\lambda}^{*} = \frac{C_1}{\lambda^5 \left[\exp\left(C_2 / \lambda T \right) - 1 \right]} \tag{5.1}$$

where we use the asterisk * to refer to black body radiation. λ is the wavelength in metres, T is the absolute temperature in Kelvin (K), C_1 = 3.74 10–16 W m^2, and C_2 = 1.44 10$^{-2\,\mathrm{m}}$ k. Planck's law shows the spectral dependence of energy emitted by a black body at different absolute temperatures (see Fig. 5.1).

The sun has an effective radiative temperature of about 6000 K, but terrestrial objects have temperatures only of the order of 300 K. At different radiative temperatures, the area under the Planck curve changes, and the spectral quality of the radiation shifts along the wavelength axis.

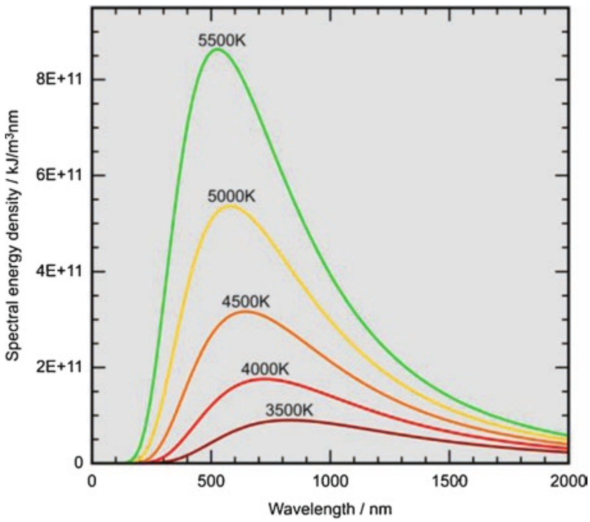


Fig. 5.1 Wien's law. (Credit: 4C [CC BY-SA 3.0 (http://creativecommons.org/licenses/by-sa/3.0/)])

Stefan-Boltzmann law. All bodies have certain temperature, emitted radiation energy. The warmer objects emit a higher energy per unit area, compared to cold objects. The total amount of radiant energy emitted by a black body is proportional to the fourth power of the absolute temperature such that:

$$E = \sigma T^4 \quad (\text{Stefan - Boltzmann law}) \tag{5.2}$$

Wien's displacement law. The wavelength of the peak in the spectral curve is given by $\lambda_{\max} = \frac{2897}{T} \mu m$, where T is the temperature in Kelvin. Wien's displacement law shows that there is a wide separation between solar radiation (shortwave) and terrestrial radiation (longwave):

Sun: $T \sim 6000$ K = > $E \sim 74.000.000$ W/cm^2
Earth: $T \sim 300$ K = > $E \sim 460$ W/cm^2

The maximum radiation emitted by warmer bodies is a shortwave length:

$$\lambda_{\max} = C / T \,(\text{Wien's Law}) \tag{5.3}$$

Sun, $\lambda_{\max} = 0.44 \mu m$; Earth, $\lambda_{\max} = 9.66 \mu m$ ($\mu m = 10^{-6} m$)

Hence, the solar radiation is called shortwave radiation, and the Earth's radiation is called longwave radiation. The objects which are good absorbers of radiation are also good emitters. A perfect absorber/emitter is called an absolute black body. The Sun and the Earth absorb/emit approximately 100% and behave like absolute black bodies. However, gases are selective absorbers/radiant, i.e. they only absorb or radiate at certain wavelengths. So, the atmosphere is transparent (visible) at some wavelengths, but non-transparent (invisible) at others. Under the influence of the atmosphere, or under the influence of the physical properties of gases that make the air flow under the influence of the suspended particles, solar radiation touches the Earth's surface, and it weakens accordingly. This represents a direct, diffuse, and global solar radiation.

5.3 Electromagnetic Radiation

The energy of the radiation travels through space in a series of electromagnetic waves. Electromagnetic radiation (EM radiation) is defined as a form of energy that is produced by the movement of electrically charged particles traveling through a matter or vacuum or by oscillating magnetic and electric disturbance. The magnetic and the electric fields (Fig. 5.2) each other come at 90°, and the combined waves move perpendicular to both electric and magnetic oscillating fields, causing the disturbance. All these rays belong to the spectrum of electromagnetic radiation (Fig. 5.3). The spectrum of electromagnetic waves is characterized by wavelength,

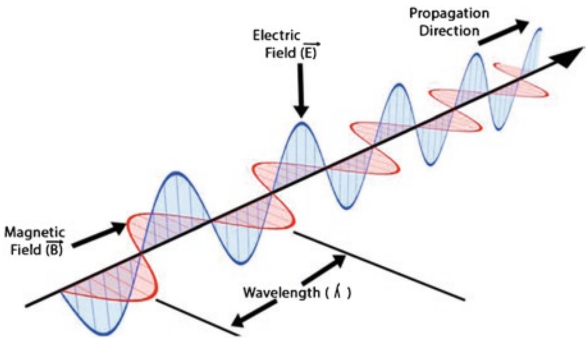


Fig. 5.2 Electromagnetic wave. (Credit: DECHAMMAKL [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

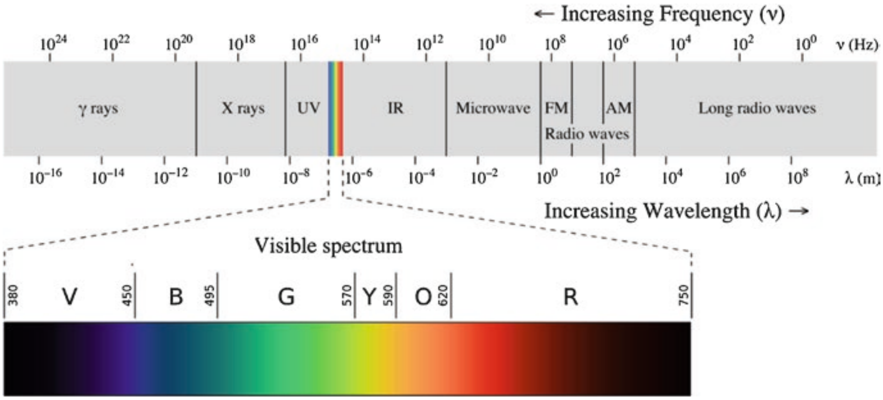


Fig. 5.3 Electromagnetic spectrum. (Credit: Philip Ronan, Gringer [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)])

from radio waves (103 m) to gamma rays (10–14 m). The Sun contains all forms of radiation, but in different amounts.

The main part of the visible spectrum ranges from ultraviolet to infrared radiation. Electromagnetic radiation means distribution of the solar energy at different wavelengths. High-energy waves are short (billion part of the spectrum and less), while low-energy waves are long waves (metres or kilometres). Shortest wavelengths are called gamma rays, while slightly longer ones are called X-rays. Out of these, visible light can be felt through the eyes, and it can even be divided into smaller parts: violet, blue, green, yellow, and red. In this array of colours, purple has the shortest wavelength, while red is characterized by the longest wavelength, if stating shortest wavelength is considered. These are the colours of the rainbow. All wavelengths shorter than infrared are called shortwaves.

5.4 Solar Radiation

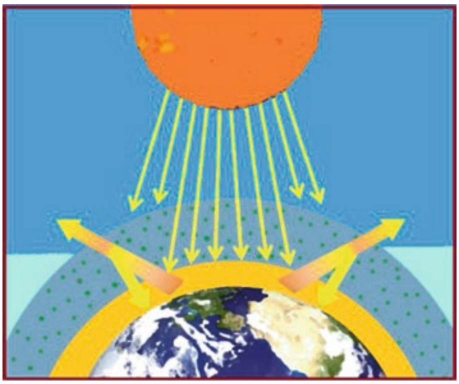
The energy released during atomic Sun activities is emitted through radiation, and often this energy is called the solar energy (Thomas and Stamnes 2006; Zdunkowski et al. 2007). This radiation has a very complex composition. Quite often in our daily life, once thinking of the solar radiation means the sunlight, but unfortunately, such an understanding is quite wrong, since sunlight is just one of the several types of energy contained in the solar energy. For example, our body exposed to this radiation primarily senses its thermal effect, from which it is concluded that it is the carrier of thermal energy.

If we are exposed to the effects of the solar radiation for a long time, some changes (e.g. redness or darkness) will occur on our skin, which are caused by energy that has a chemical effect. Solar radiation is radiant energy or particularly electromagnetic energy emitted by the Sun that propagates in all directions in space (Fig. 5.4). About half of the radiation belongs to the visible shortwave part of the electromagnetic spectrum. The other half is mostly in the near-infrared part, with some in the ultraviolet part of the spectrum. From the total emitted energy by the Sun, approximately billion parts come at the Earth's surface, which is quite enough to warm the air and the temperature necessary for living. Under the influence of the Sun, all processes and phenomena in the Earth's atmosphere are driven.

5.4.1 Solar Constant

Earth receives just one small portion of the solar energy. We can quantitatively measure such energy, only in the upper part of the atmosphere. To determine the portion of solar radiation that participates in atmospheric processes, it is necessary to determine the amount of solar radiation that comes to the Earth's surface. But the solar radiation is strongly influenced by the atmosphere. Therefore, the total solar radiation can be determined in the upper part of the atmosphere or at the Earth's surface,

Fig. 5.4 Solar radiation



but without the influence of the atmosphere and assuming the average distance of the Earth from the Sun. The total radiation energy received from the Sun per unit of time per unit of area on a theoretical surface perpendicular to the Sun's rays and at Earth's mean distance from the Sun is called solar constant. The solar constant is not a true constant because the change of 2–5% depends on the distance between the Sun and the Earth in the annual seasons, spring, summer, autumn, and winter, due to the appearance of Sun spots and perturbations. In another term, solar constant is the amount of energy potentially available for the Earth. Insolation is the amount of solar radiation received on a given surface in each time period. Daily insolation is the incident solar radiation on a horizontal surface per square metre integrated over a day. Insolation varies seasonally and with daily variation in cloud cover. Namely, there is always a difference between the solar constant and insolation, because insolation, i.e. energy received on Earth, is less than 50%. This means that there is a loss of about 50% of solar energy in the atmosphere.

5.4.2 Direct Solar Radiation

The solar radiation that covers our sky can be direct, diffused, or reflected radiation (Fig. 5.5). The Sun always emits brightly when direct sunlight is falling on the Earth's surface. The sunlight that reaches the Earth's surface unmodified by any of the atmospheric processes is called direct or beam solar radiation. The intensity of direct solar radiation depends on the height of the Sun above the horizon, the height above sea level, cloudiness, transparency of the sky, the amount of water vapour in

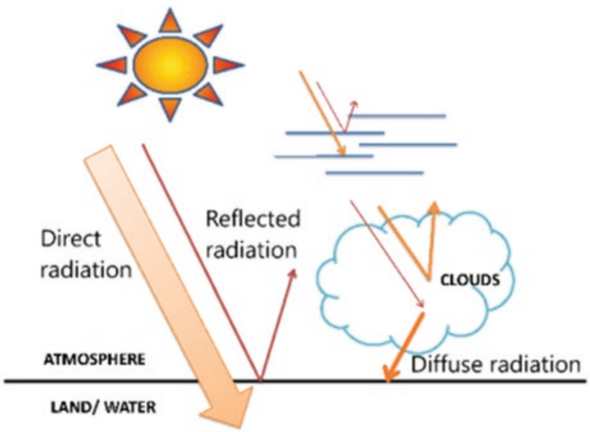


Fig. 5.5 Direct, reflected, and diffused radiation

the air, and other factors. The sum of direct solar radiation (S) and diffused radiation (D) gives the global radiation (G):

$$G = S + D$$

The intensity of global radiation depends on the same factors as the intensity of direct radiation and it diffuses. Sunlight is falling on the Earth's surface from a different angle and a different direction. Thus, it causes a different heat effect. For example, the open solar radiation lasts all day long, from sunrise to sunset, and the Sun reaches its maximum value in the afternoon. During the summer, solar radiation has the smallest value in the afternoon and also in winter when the Sun has its lowest position in relation to the horizon. Insolation value changes with the changing of the angle of inclination of the Earth's surface and changing of the geographic latitude (Fig. 5.6).

5.4.3 Diffused Solar Radiation (D)

In addition to direct incoming solar radiation, there is also diffused radiation. It usually occurs by diffusive reflection of direct solar radiation, when it passes through the atmosphere. In a collision with molecules of different gases, water vapour, tiny mineral, or organic compounds present in the atmosphere, sunlight is reflected in different directions. The phenomenon, in which the Sun's rays change its direction and wavelength, is called diffuse reflexivity. This type of radiation appears under the clear sky conditions at the shortwave rays: blue, blue, violet, and ultraviolet rays.

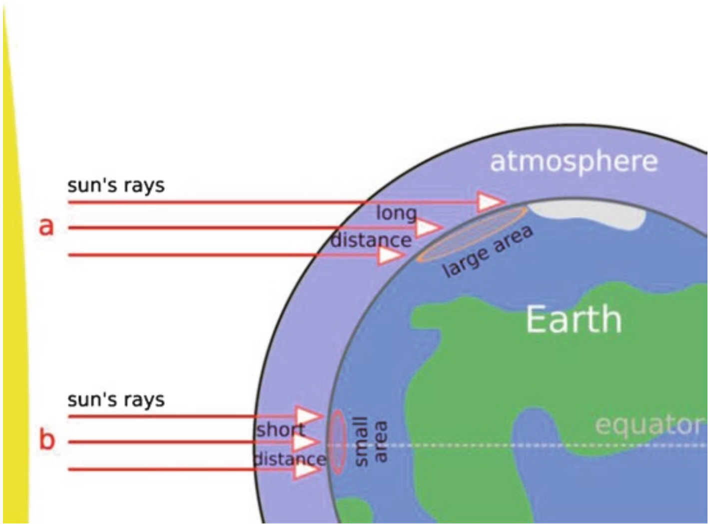


Fig. 5.6 Effect of the Earth's shape and atmosphere on incoming solar radiation. (Credit: Peter Halasz |CC-BY-SA-2.5| https://creativecommons.org/licenses/by-sa/2.5/deed.en)

Diffused radiation lasts from the moment of sunrise to the time of sunset. Diffused intensity of radiation depends on the height above the sea at the place where it appears from cloudiness, transparency of the atmosphere, and the location of the Sun in the sky.

5.4.4 Solar Radiation Factors

Absorption. The atmosphere interacts with both incoming solar radiation and outgoing terrestrial radiation. The strength of the interaction as a function of wavelength is responsible for the heating of the lower atmosphere. A process in which atmospheric gases and particles reduce the amount of solar or terrestrial radiation is called absorption. Such process takes place through the transfer of energy that occurs when increasing molecular movement. Due to the chemical composition of the Earth's atmosphere, most of the infrared radiation emitted by the warm surface never reaches space. Instead, the radiation is reflected or absorbed by compounds known as greenhouse gases. When these compounds absorb the infrared radiation from the surface, the atmosphere heats up. The energy reflected towards the Earth warms the surface further, causing the Earth to emit more infrared radiation. This creates a cycle that keeps the atmosphere and the surface warm. As it is mentioned above, this process results in the heating of the lower atmosphere, which mainly consists of gases with different absorptivities. For example, N_2 is a poor absorber, but O_2, O_3, and H_2O are effective absorbers, and they are responsible for most of the absorption in the atmosphere.

Reflection. Reflection is a process of transfer of solar radiation at the surface of the Earth. Reflected sunlight depends from albedo and incoming sunlight.

Albedo. Albedo is a measure of the ability of the surface to reflect solar radiation, i.e. if it is a reflector or albedo = 1, it is an ideal reflector. In Table 5.1, albedo values for different surfaces of the Earth are given.

The highest reflection of solar radiation (albedo) occurs in cases of freshly fallen snow (75–95%), while the smallest value of albedo (5–10%) is noticed in dark

Table 5.1 Reflection of solar radiation: albedo

Object or surface	Albedo (%)
Fresh snow	75–95
Old snow	40–60
Sand	20–30
Ground	15–25
Black street	5–10
Thick cloud	70–80
Thin cloud	25–30

backgrounds of the Earth's surface (dark streets, bases, land, etc.). Albedo also changes the angle of the Sun: albedo water changes from 5% at high Sun to 80% in the occurrence of low Sun. The overall albedo of the Earth and the atmosphere is approximately 30%, with the largest percentage of presence of clouds and less than sea-land areas.

Scattering. Scattering means redirection of incoming solar radiation in all directions, not just back to space. Solar radiation travels through a straight line, and then it is redirected (scattered) by gases (Rayleigh) or by aerosols (Mie). Scattering changes the direction, but does not change the wavelength of light. Thereby, gases and aerosols are more effective in scattering radiation with different wavelengths. This causes a blue sky.

Rayleigh scattering is notable when the radius of the particles is about 1/10 times the order of wavelength of electromagnetic radiation illustrated in the molecules of the air and can be extended to scattering from particles up to about a tenth of the wavelength of the light. It is Rayleigh scattering off the molecules of the air which gives us the blue sky.

Mie scattering occurs when the particles have a larger diameter than electromagnetic radiation. It is caused by pollen, dust, smoke, water droplets, and other particles in the lower portion of the atmosphere. It occurs when the particles causing the scattering are larger than the wavelengths of radiation. Figure 5.7 shows a graphical depiction of the three mechanisms of scattering:

1. Rayleigh scattering
2. Non-selective scattering (wavelength independent)
3. Mie scattering

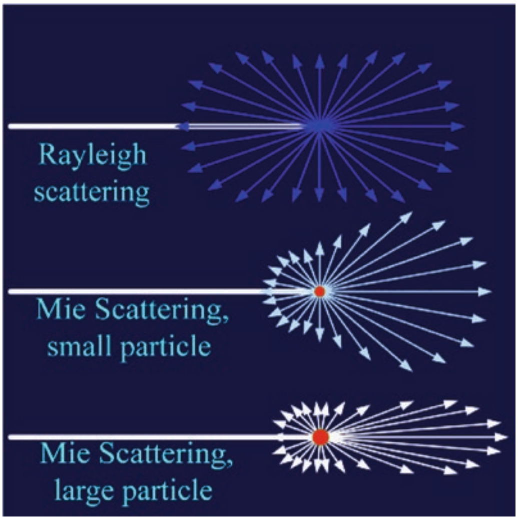


Fig. 5.7 Graphical illustration of three types of scattering

5.4.5 Temporal and Spatial Changes in Insolation

Earth's energy balance refers to an arbitrarily 100 units of radiation taken. The amount of insolation is not constant in time and space. Insolation changes with the change of season and the change of geographical latitude. This change of incoming solar radiation is manifested as latitudinal and seasonal change in temperature (T).

To understand these changes, it is necessary to know certain physical laws and relations in the Earth-Sun system.

Earth's movement. Earth's rotation around its own axis affects the changes daily. The movement of the Earth around the Sun is called *revolution.* The distance between the Earth and the Sun changes, but it has a little role in seasonal changes. For example, the Land of the Northern Hemisphere is closest to the Sun during winter.

Seasonal and latitudinal change of insolation. An important factor for the seasonal changes in latitudinal insolation represents the height of the Sun (i.e. the angle that affects sunlight to the horizon). In summer, the midday Sun is high in the sky, but in winter it is lower. The altitude of the Sun affects the energy that arrives at the Earth's surface. When the Sun is at a lower angle in relation to the horizon, the Sun rays pass through most of the atmosphere, and escaping radiation has a greater opportunity to absorb or reflect. The country has a spherical shape, so that the vertical rays arrive only in places with a certain latitude (i.e. when the sun is at 90°, the zenith). Moving north or south, the angle of the Sun and the length of the day are changing. From here the amount of Sun, solar energy, and therefore the Earth's surface change with geographical latitude. The altitude of the Sun changes with the season. The Earth's axis is at an angle of 23.5° with flat orbit around the Sun (so-called angle of inclination). In this case, the Earth's orientation to the Sun changes with time and here in width. For example, in June, the Northern Hemisphere is leaning on the side of the Sun, while the Southern Hemisphere is further away from it. In December, the Northern Hemisphere is leaning towards the Southern Hemisphere away from the Sun. Each year the 4 days are given special importance owing to the annual shift of the direct rays of the Sun. These specific days are of importance to the annual cycle of time and are shown in Fig. 5.8. Latitudes 23.5° N and 23.5° S are known as tropics of Cancer and Capricorn, respectively. During summer solstice, geographical latitudes of the Northern Hemisphere have longer days than the Southern Hemisphere. The area above 66.5° N (Arctic Circle) has a constant daylight, while below 66.5° N (Antarctic Circle) has a permanent night. Fully opposite picture appears during the winter solstice. This resulted in the occurrence of higher temperatures in summer than in winter. Seasonal changes in the amount of solar energy are caused by migration of the vertical rays of the Sun, resulting in changes in the angle of the Sun and the length of the day.

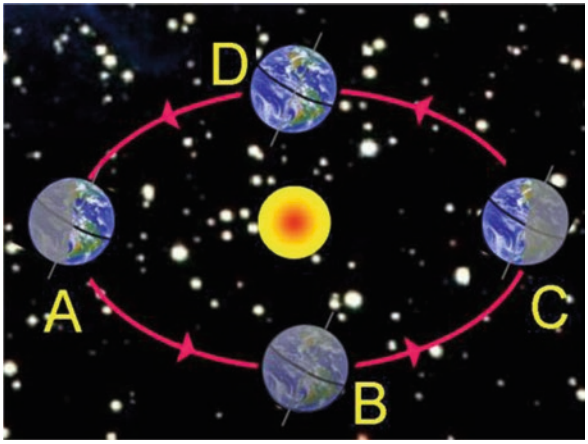


Fig. 5.8 Earth's revolution: equinox, solstice, seasons, perihelion, and aphelion

5.5 Optical Radiation

Optical radiation is a part of electromagnetic spectrum that covers wavelengths ranging from 100 to 10,000 nm. This range of wavelengths can be divided into three areas:

- (100–400 nm) – ultraviolet (UV) region
- (400–770 nm) – visible (VV) field
- (770–10,000 nm) – infrared (IR) area

The *ultraviolet part* of the spectrum is the radiation with the shortest wavelengths that are not visible to the human eye, but with very high energy and harmful action. Radiation from the visible part of the spectrum can be registered (it is visible) by the human eye. This part of the spectrum consisted of six components with different colours, purple, blue, green, yellow, orange, and red, arranged from the shortest to the longest wavelength. The length of radiation waves belonging to the infrared part of the optical spectrum is emitted by intensively warmed bodies, so this part of the spectrum is often called heat radiation. Infrared radiation does not have a strongly detrimental effect, which means that with increasing wavelength, the harmful solar radiation is reduced. *Ultraviolet radiation (UV).* The ultraviolet radiation in the optical part of the electromagnetic spectrum (Fig. 5.9) is covering an area with the shortest wavelengths (see Vasquez and Hanslmeier 2006). There are three different types of ultraviolet (UV) radiation, based on the wavelength of the radiation and according to the biological activity and the effects it causes:

- UV-C radiation (100–290 nm)
- UV-B radiation (290–320 nm)
- UV-A radiation (320–400 nm)

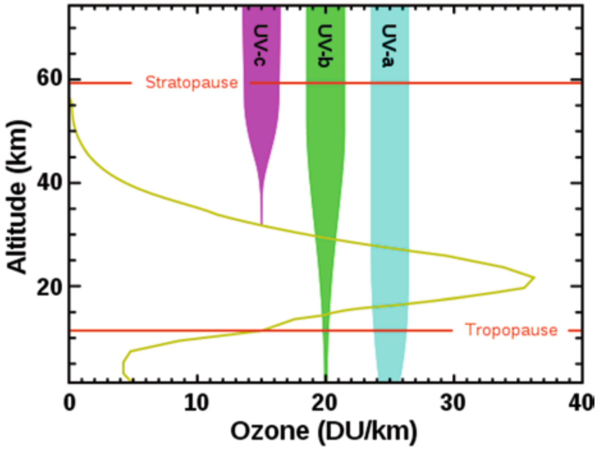


Fig. 5.9 Ozone levels at various altitudes (DU/km) and related blocking of several types of ultraviolet radiation

UV-C radiation is completely absorbed in the upper atmosphere by the oxygen and the ozone molecules. Most of the UV-B radiation is absorbed in the stratosphere by ozone molecules, and only part of the radiation is coming to the Earth's surface.

Given that ozone is the main absorber of UV-B radiation, the intensity of this radiation at the Earth's surface much depends on the total amount of ozone in the atmosphere and beyond, from its amount in the depths of the ozone layer. It is known that UV-B radiation is biologically very harmful. On the surface of the Earth, the UV-A radiation comes in the largest amount. UV-A radiation is less harmful, but it is possible to cause a dark tan skin in humans.

If the amount of UV is high enough, the ability of self-protection of individual biological units is not enough, and the individual may be damaged. It refers to the human body, especially the skin and the eyes.

5.5.1 UV Index

To avoid damage when exposed to UV, the UV index should be made public in order to warn and inform people about the harmfulness of UV and the ways to take certain precautions. UV radiation can be measured as radiation or irradiation energy received at unit area in units (W/m^2) or radiation (radiating exposure) or dose of energy per unit area summarized in a specific time interval in units (W/m^2). The global solar UV index (UVI) shown in Fig. 5.10 describes the level of solar UV radiation at the Earth's surface. The values of the index range from zero upwards – the higher the index value, the greater the potential for damage to the skin and eye and the less time it takes for harm to occur.

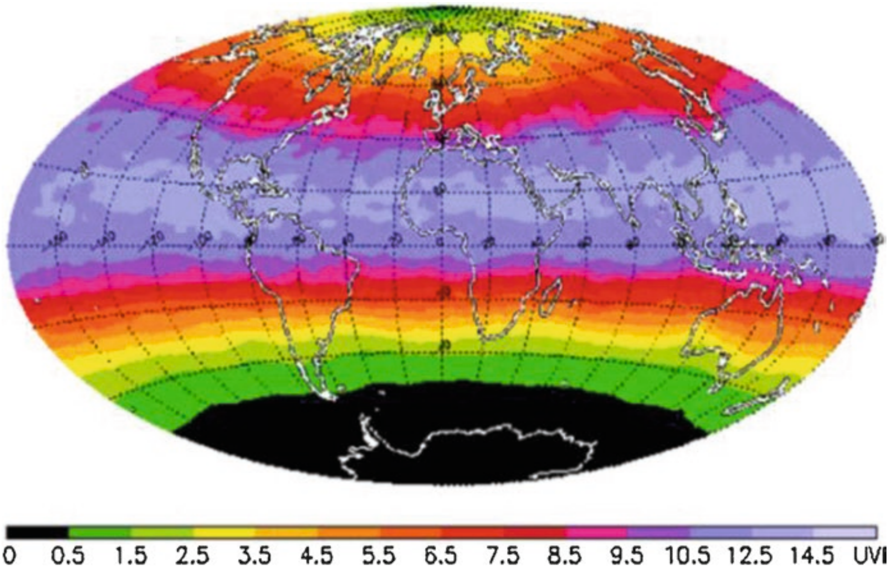


Fig. 5.10 Global UV index distribution

5.5.2 UV Index Factors

The intensity of UV radiation depends on the following factors: the amount of Sun, altitude, atmospheric scattering, cloudiness, and reflection from the ground (soil).

Height of the Sun. It represents a Sun elevation angle below which the amount of Sun in relation to the direction of horizon can be seen. The solar zenith angle is often used instead of the amount of the Sun. It is the zenith angle and direction of the Sun. Elevation has great value when the Sun is high, because the way through the absorption layer of the atmosphere is the longest and then most photons come to the surface of the Earth. Solar elevation is highest in the middle of the day in summer. At the equator, the angle of the sunlight that falls on the Earth's surface has the greatest value. In this case, the intensity of UV in places with 0-degree latitude will be larger in terms of areas with higher latitudes. Equality intensity of UV is the greatest since the value of elevation is minimal. UV is strongest in tropical areas, in summer at noon.

Elevation height. UV increases with height above the sea, because of the smaller number of absorption components in the atmosphere. Measurements show that UV increases by 6–8%, with increasing height above the sea from 1000 metres.

Atmospheric scattering. Solar radiation, which goes to the Earth's surface, is divided into direct and diffused. On its way to the Earth, the radiation is scattered by molecules of gas and other particles present in the atmosphere: aerosols, water droplets, particles, and dust (Coakley and Yang 2014). Direct radiation is the radiation which passes directly through the atmosphere, without scattering or absorption.

Diffused radiation is the radiation that is, at least once, being scattered in the atmosphere before reaching the ground. Scattering depends on the wavelength. The sky looks blue because blue light is much stronger and scattered than other wavelengths of visible light. UV radiation is scattered even more, so that the Earth's surface UV-B radiation is comprised of approximately equal scale (1:1), of direct and diffused radiation.

Clouds and atmospheric haze. UV is more intense in the open sky without clouds. Clouds basically reduce UV radiation. The type of the clouds and their thickness changes and impacts UV. Partial cloudiness (presence of thin or broken clouds) has only little influence of UV on the Earth's surface. In certain conditions and for a short period of time, small cloudiness at some places may even enhance UV radiation, compared to a completely open sky. In terms of atmospheric haze, UV is absorbed by water and dispersed particles and aerosols, and this leads to a reduction in UV.

Reflection of the Earth's surface. One part of UV that reaches the Earth's surface is absorbed by the Earth's surface, while another part is reflected to space. Reflected amount of radiation depends on the properties of the Earth's surface. Most natural surfaces, such as grass, ground, and water, reflect less than 10% of the received UV radiation. Fresh snow can reflect up to 80% of the received UV radiation. During the spring, in the open sky, reflection in the snow can increase UV radiation of declined surfaces, even to the summer values. It is important for the heights above the higher latitudes. Sand can reflect about 25% of UV and can increase the UV exposure of the beach. Up to 95% of UV can penetrate the water and about 50% can penetrate up to 3 m in depth (in the clean oceanic water). The intensity of the UV increases to 6% for each kilometre altitude. Sky without clouds passes ~100% (UV-C), low clouds ~89%, middle clouds ~73%, and complete cloudiness around 31% of UV.

5.6 Ozone Layer

Ozone is very rare in the atmosphere. Every ten million molecules of air are used for about three molecules of ozone. Despite this small amount, ozone plays a crucial role in the atmosphere. Ozone is mainly located in two regions of the atmosphere. Approximately 90% is in the ozone layer that begins at 10 and 17 km above the Earth's surface and extends to about 50 km. This region of the atmosphere is called stratosphere. This layer is known as the ozone layer or ozonosphere, because of the presence of ozone.

Residual ozone can be found in the lower part of the atmosphere, which is popularly called troposphere. Molecules of ozone in the upper atmosphere (stratosphere) and the lower layer of the atmosphere (troposphere) are chemically identical, because they all consist of three atoms of oxygen and have the same chemical formula, O_3. However, they have very different roles in the atmosphere and very different effects on humans and on other living beings. Stratospheric ozone has a beneficial role, because it absorbs most of the biologically harmful solar ultraviolet (UV-B), allowing only a small amount to reach the surface of the Earth. Absorption of ultraviolet radiation by ozone creates a source of heat, which forms the stratosphere (region in which temperature increases with height). Ozone also plays a key role in the temperature structure of the Earth's atmosphere. Since there is no protective effect of the ozone layer, a greater amount of sunlight (UV-B) radiation enters the atmosphere and reaches the surface of the Earth. With the help of many experimental studies of plants and animals, as well as through clinical studies on humans, the harmful effects of excessive exposure to (UV-B) radiation have been shown. The dual role of ozone leads to two separate environmental issues. Near the Earth's surface, ozone is a key component of photochemical "smog", a well-known problem in the atmosphere of many cities in the world, but there is a trend of increasing surface ozone in rural areas. Also, there is a widespread scientific and public interest and concern over the reduction of ozone in the stratosphere and the appearance of the ozone hole. The amount of reduction of stratospheric ozone in the atmosphere is measured using modern instruments and satellites. Over some parts of Antarctica, up to 60% of the total amount of ozone which lies in the column was exhausted during spring (September to November). This phenomenon is known as the Antarctic ozone hole. A NASA instrument has detected an Antarctic ozone "hole" or ozone depletion region expanded to largest dimension of about 11 million square miles on September 3, 2000 (Fig. 5.11). The minimum amount of ozone for about 85 Dobson units is registered on October 8, 2006. However, the ozone hole's size currently has stabilized, but the low levels in its interior continue to fall. Based on scientific evidence obtained in the last two decades, certain chemicals produced by man are responsible for the depletion of the ozone layer. Compounds that deplete the ozone layer contain various combinations of chemical elements like chlorine, fluorine, bromine, carbon, hydrogen, and so-called halocarbons. Compounds containing only chlorine, fluorine, and carbon are called chlorofluorocarbons, usually abbreviated as CFCs. Chlorofluorocarbons, carbon tetrachloride, chloroform, and methyl are important gases produced by humans, which affect and cause harmful ozone depletion, and are used in many applications, including refrigeration, air conditioning, cleaning of electronic components, as well as solvents.

5.7 Earth Longwave Radiation

While the Sun emits energy in the shortwave spectrum, the Earth emits energy from the surface to the atmosphere in a longer wavelength (Herbin and Dubuisson 2015). This longwave Earth's radiation is known as *terrestrial radiation* (TR). The Earth's surface is heated to 20 °C approximately on average by the absorption of solar radiation. As a result, it emits backward thermal infrared radiation (IR) with wavelength of 5–50 μm up to 50 μm. There is so-called atmospheric window in which band the atmosphere is transparent, allowing infrared radiation to escape to space. On both sides of this narrow band, there are absorption bands due to the presence of atmospheric gases.

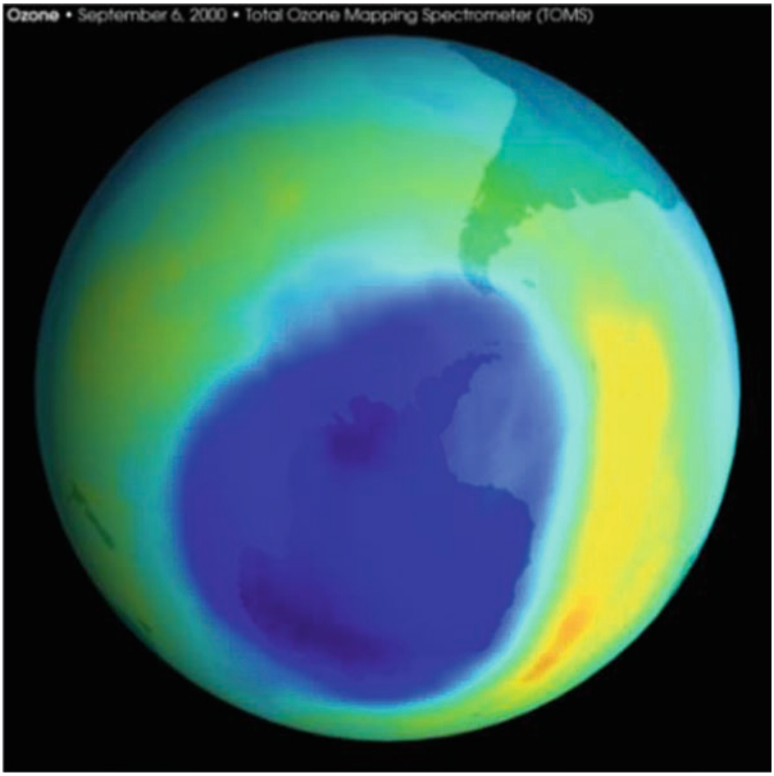


Fig. 5.11 Ozone hole. (Image of the largest area of Antarctic ozone thinning ever recorded in September 2000. Data taken by the Total Ozone Mapping Spectrometer (TOMS) instrument aboard NASA's Earth Probe satellite)

The distribution of 1979–1995 global mean outgoing longwave radiation in W/m^2 is shown in Fig. 5.12. While the atmosphere is only a weak absorber of solar radiation (insolation), it is an effective absorber of terrestrial radiation or Earth longwave radiation. It means that gases within the atmosphere absorb wavelengths in longwave radiation spectrum of the Earth, unlike the solar radiation. The amount of solar radiation reaching the Earth's surface depends on the latitude, season, and time of day, cloudiness, aerosol backscatter, surface topography, and its roughness. Water vapour (H_2O) and carbon dioxide (CO_2) are the main absorbers; ozone (O_3), methane (CH_4), and other gases also have absorption abilities; and H_2O absorbs five times more than other gases. The reason for the high temperatures (T) in the lower troposphere, and the decline in air temperature with height (the atmosphere is heated from the ground floor layer up), is because the atmosphere is transparent to insolation, but the Earth absorbs radiation. When the gases absorb the Earth's radiation, they are heated and emit energy in all directions. Part of this energy is returned to the Earth's surface and further heated. This is called "greenhouse effect". Gases that absorb the Earth's longwave radiation (TR) are called "greenhouse gases". The total

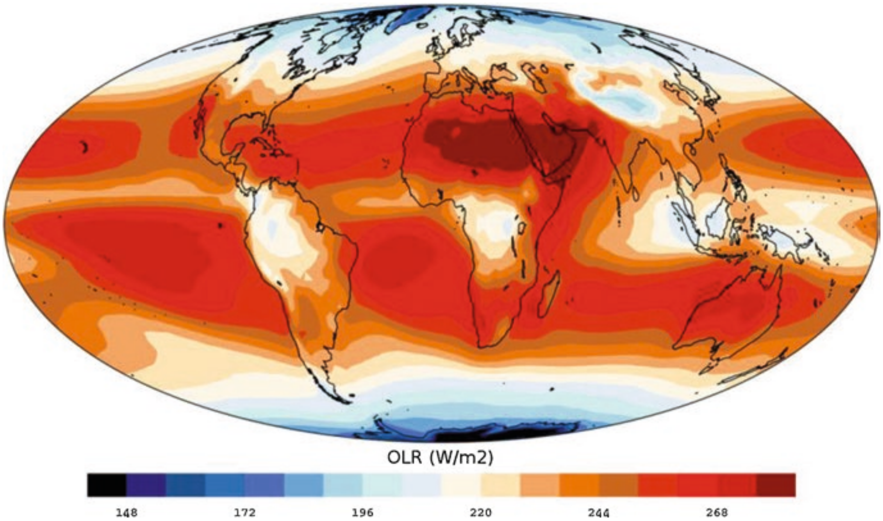


Fig. 5.12 Outgoing longwave radiation. (Credit: Giorgiogp2 [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)])

radiation of the Earth may be a different energy, i.e. with a different frequency or wavelength. What is common to all forms of electromagnetic radiation is their nature. Radiation at all wavelengths together constitutes the electromagnetic spectrum. All bodies warmed to a temperature emit electromagnetic radiation. At temperatures typical of the atmosphere, bodies emit radiation of wavelengths from the infrared part of the spectrum, while at higher temperatures, bodies emit radiation from the visible part of the spectrum. Emitted radiation of these bodies occurs because of transformation of their internal energies in the electromagnetic radiation, and the spectrum of radiation depends on the temperature to which the body is heated. The distribution of electromagnetic radiation that is emitted from the surface of the Sun and which touches the upper layer of the atmosphere, a function of wavelength, is called solar spectrum.

5.7.1 Earth's Annual and Global Mean Energy Budget

The Earth's energy budget accounts for the balance between the energy the Earth receives from the Sun and the energy the Earth radiates back into outer space after having been distributed throughout the five components of the Earth's climate system and having thus powered the so-called Earth's heat engine (Melnikova and Vasilyev 2005; Hewitt and Jackson 2003). Partitioning of the radiative energy throughout the atmosphere is achieved using detailed radiation models for both the longwave and shortwave spectral regions. Here is a great chart (Fig. 5.13) that very nicely shows the key heat flows that determine the Earth's surface temperature, the

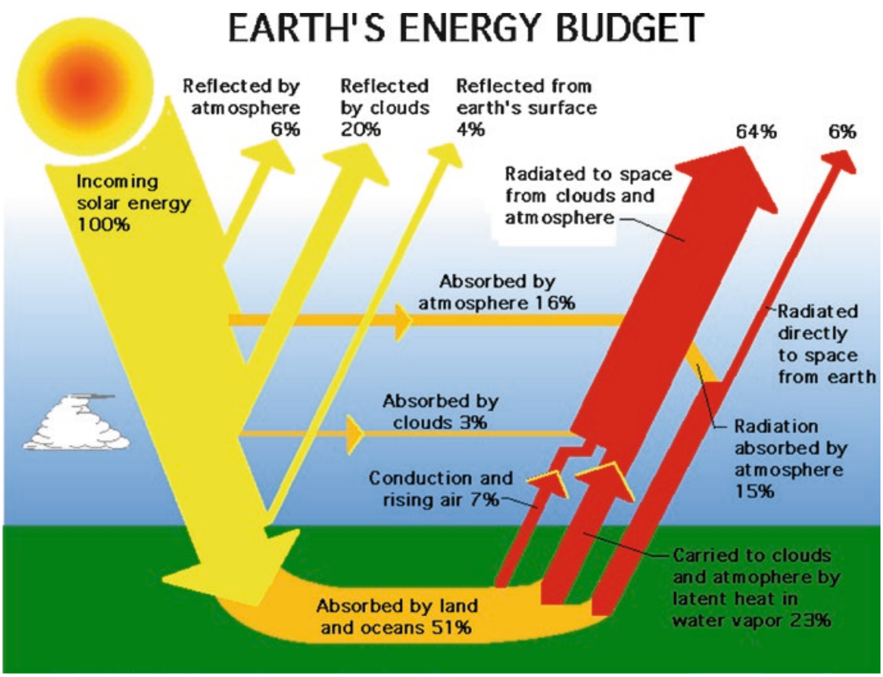


Fig. 5.13 Earth's energy budget

Earth's energy budget diagram. Incoming sunlight is on the left; outgoing infrared or "longwave" radiation is on the right. The values illustrate the relative magnitude of each term, contributing in the energy budget. The following is a description of a simple model based on the energy budget of radiation. It contains the basic elements of energy fluxes which depend on climate. Values in squares are inflows of energy or flux in units of watts per square metre (W m^{-2}). The energy that is emitted by the Sun, taking into account the averaged Earth's surface, is 342 W m^{-2}. Out of this amount of energy, 77 units are reflected back and thrown into space from the atmosphere, including clouds and aerosols. The clouds are a special case of aerosols, i.e. particular matters with dimensions much larger than those of the molecules.

Other aerosols include drops of sulphuric acid formed by oxidation in the atmosphere of volatile sulphur compounds. With the passage of solar radiation through the atmosphere, a different amount of energy (67 W cm^{-2}) is absorbed by greenhouse gases. Ten of these 67 units absorbed in the stratosphere, mainly from molecules of oxygen, under the influence of high-energy ultraviolet photons are split into two oxygen atoms. Reactive atoms of oxygen unite with oxygen molecules $O_2 \rightarrow 2O$ when they get ozone: $O + O_2 \rightarrow O_3$. This reaction is vital for life, because ozone absorbs most of incoming ultraviolet radiation, the remaining part of the solar radiation which is transferred from the atmosphere, undergoing yet another reflection of the amount of surface area of 30 W cm^{-2}, and only 168 W cm^{-2} is absorbed by the Earth's surface. Consequently, the surface is heated by infrared radiation which is sent downwards by the atmosphere. The Earth's energy budget shows the distribution of incoming radiation received from the Sun and the outgoing radiation which is emitted by the Earth over the entire planet Earth. It is obvious that there is some distortion of the balance in latitudinal distribution of radiation. The energy shortage exists in the northern latitudes of both hemispheres. This leads to the cooling of the atmosphere, while the excess energy that exists in the equatorial area implies heating of the atmosphere. As a result of differences in the latitudinal distribution of energies, transfers (transfer) of heat from the equatorial towards the polar regions appear.

5.7.2 Earth's Heat Balance

The annual balance that exists between incoming and outcoming radiation, called the Earth's heat balance, contributes to the mean temperature on the Earth and is maintained relatively constant despite seasonal penetration of cold and warm waves. Although the balance between incoming and outgoing radiation is maintained throughout the planet, it does not happen at every latitude. During the year, the belt on the Earth's surface between 36° N and 36° S latitude, on average, receives a larger amount of solar radiation than it loses in space. Conversely, at higher latitudes, greater amount of radiation is left by the longwave Earth's radiation compared to the amount that the Earth receives. As a result of energy imbalance between the low and high latitudes, global winds and ocean currents raise, which is a way to transfer excess heat from the tropics towards the poles. Tables 5–7 give a schematic representation of the overall energy budget for the Earth's surface, atmosphere, and the entire planet. Tables are global annual means expressed as a percentage of the rate per unit area at which solar radiation is intercepted by the Earth; i.e. 100 units is equivalent to 342 W m^{-2}. Table 5.2 clearly illustrates that the Sun arrives directly on the surface of the Earth, unit 51, while the atmosphere radiates 95 units, or a total of 146 units. Earth longwave radiation is 116 units. Through evaporation and convection from the Earth's surface, 30 units or 146 units in total are released, so obviously there is a balance between incoming and outgoing radiation of heat in the Earth's surface. The heat budget of the atmosphere is shown in Table 5.3. Most of the incoming radiation is received by the atmosphere from the Earth's longwave radiation of 110 units, and only 19 units are received from direct solar radiation.

Table 5.2 The heat budget of the Earth's surface

Input		Output	
Solar radiation	51	Earth radiation	116
Atmospheric radiation	95	Evaporation	23
		Convection	7
	146		**146**

Table 5.3 Heat budget in the atmosphere

Input		Output	
Solar radiation	19	Radiation to space	64
Condensation	23	Radiation to the surface	95
Convection	7		
Earth radiation	110		
	159		**159**

Table 5.4 Heat budget of the entire planet

Input		Output	
Solar radiation	100	Reflected/scattered radiation from clouds or gases	30
		Atmospheric radiation to space	64
		Earth to space longwave radiation	6
	100		**100**

Most of the output radiation from the atmosphere is directed towards the Earth's surface. It is evident that there is a balance between incoming and outgoing radiation of the atmosphere, in which there are 159 units of input radiation in relation to 159 units of output radiation.

Heat budget of the entire planet (Table 5.4) is considered as a balance between incoming shortwave radiation of the Sun, which is 100 units, and outgoing radiation, which consists of the Earth's longwave radiation to space of 6 units, atmospheric radiation to space of 64 units and reflected or scattered radiation from clouds or gases of 30 units or a total of 100 units.

It is obvious that there is some distortion of the balance in latitudinal distribution of radiation. The energy shortage exists in the northern latitudes of both hemispheres. This leads to the cooling of the atmosphere, while the excess energy that exists in the equatorial area implies heating of the atmosphere. As a result of differences in the latitudinal distribution of energies, transfers (transfer) of heat from the equatorial towards the polar regions appear.

5.7.3 Earth Radiation Budget and the Planetary Temperature

As it is described in the previous topic, the flux of solar energy arriving at the Earth (see Fig. 5.14) is called the solar constant:

$$S_0 = \frac{Q}{4\pi r^2} = 1367\,\mathrm{W\,m^{-2}}$$

The amount of radiative energy a body emits and the wavelength at which it is emitted, according to Wien's law, depend on temperature. The total radiation energy

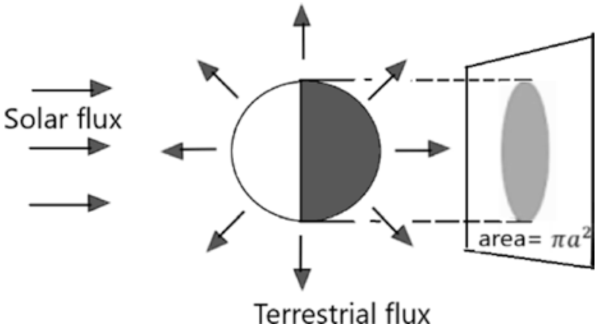


Fig. 5.14 Solar and terrestrial flux

emitted is σT^4 where T is the absolute temperature and σ is Stefan-Boltzmann constant is 5.67×10^{-8}Wm^{-2}K^{-4}.

Hence, the incoming shortwave solar radiation is:

$$E_{\text{solar}} = S_0 \pi a^2 \tag{5.4}$$

If the fraction of incident solar energy that is reflected is denoted as (α), which represents albedo (planetary albedo = 0.3), the absorbed shortwave solar radiation is:

$$E_{\text{solar\_absorbed}} = (1-\alpha) S_0 \pi a^2 \tag{5.5}$$

If the Earth emits like a black body of uniform temperature, T_e, then the emitted radiation per unit area is given as:

$$E_{\text{earth}} = \sigma T_e^4 \tag{5.6}$$

Or the total emitted radiation by the Earth is:

$$E_{\text{earth\_tot}} = 4\pi a^2 \sigma T_e^4 \tag{5.7}$$

Since the absorbed solar radiation is equal to emitted radiation, it follows that the planetary emission temperature is:

$$T_e = \left[\frac{S_0 (1-\alpha)}{4\sigma} \right]^{\frac{1}{4}} \approx 255\text{K} \tag{5.8}$$

Atmosphere is transparent in the visible, at the peak of the solar spectrum. It is non-transparent in the UV spectrum and with variable transparency across the infrared spectrum (i.e. it is completely opaque at some wavelengths, yet completely transparent at others). Figure 5.15 shows the absorption bands in the Earth's

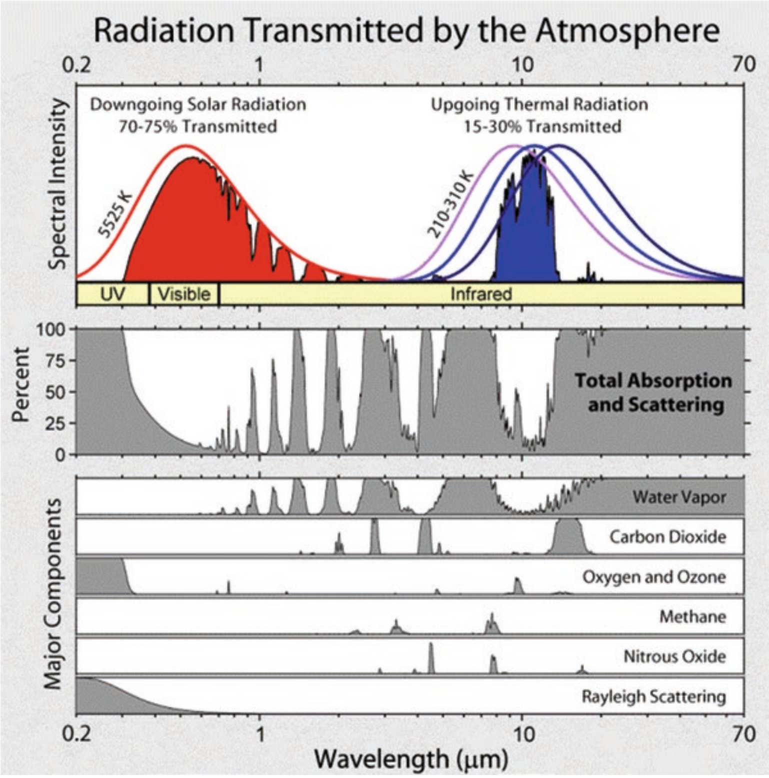


Fig. 5.15 Radiation transmitted by the atmosphere and the absorption in the Earth's atmosphere. (Credit: Robert A. Rohde for the Global Warming Art project https://creativecommons.org/licenses/by-sa/3.0/deed.en)

atmosphere (middle panel) and the effect that this has on both incoming solar radiation and outgoing thermal radiation (top panel).

The absorption spectrum for major greenhouse gases including the Rayleigh scattering is depicted in the lower panel. Both the Earth and the Sun emit electromagnetic radiation (e.g. light) that closely follows a black body spectrum. For the Sun, these emissions peak in the visible region and correspond to a temperature of ~5500 K. Emissions from the Earth vary following variations in temperature across different locations and altitudes, but always peak in the infrared. The position and number of absorption bands are determined by the chemical properties of the gases present. In the present atmosphere, water vapour is the most significant of these greenhouse gases, followed by carbon dioxide and various other minor greenhouse gases. In addition, Rayleigh scattering, the physical process that makes the sky blue, also disperses some incoming sunlight. Collectively, these processes capture and redistribute 25–30% of the energy in direct sunlight passing through the atmosphere. By contrast, the greenhouse gases capture 70–85% of the energy in upgoing thermal radiation emitted from the Earth's surface. Dominant components of the atmosphere are incredibly transparent across the important spectral range. Nitrogen doesn't absorb at all and oxygen absorbs only in the far UV (where there is a little solar flux). Absorption of terrestrial radiation is dominated by triatomic molecules (O_3, H_20, and CO_2) which have rotational and vibrational modes that can be easily excited by infrared radiation. These atmospheric constituents present in tiny concentrations, so-called greenhouse gases, play a significant role in the Earth-climate system.

5.7.4 The Simple Greenhouse Model

Figure 5.16 shows a simple greenhouse gas model. We start from the estimation of the averaged incoming solar energy E_{solar} per unit area:

$$E_{\mathrm{solar\_mean}} = \frac{E_{\mathrm{solar}}}{4\pi a^2} = \frac{S_0 \pi a^2}{4\pi a^2} = \frac{S_0}{4}$$

If we take into account absorptivity ($\varepsilon_{\mathrm{atm}}$) as fraction of the longwave radiation absorbed, and emissivity ($\varepsilon_{\mathrm{earth}}$) as fraction of black body radiation emitted by the Earth, then we can calculate the net radiative flux:

$$E_{\mathrm{netsolar\_top}} = \frac{S_0}{4}(1-\alpha) = (1-\varepsilon_{\mathrm{atm}})\sigma T_s^4 + \varepsilon_{\mathrm{atm}} \sigma T_a^4 \tag{5.9}$$

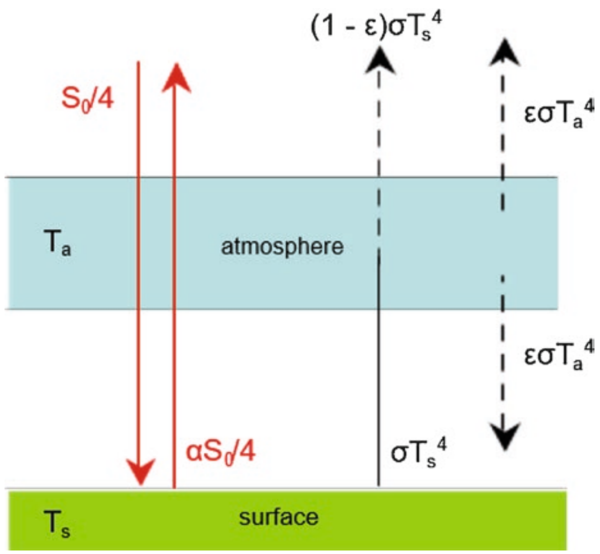


Fig. 5.16 Greenhouse gas model

(The top of the atmosphere)

$$E_{\text{netsolar\_ground}} = \frac{S_0}{4}(1-\alpha) + \varepsilon_{\text{atm}}\sigma T_a^4 = \varepsilon_{\text{earth}}\sigma T_s^4 \tag{5.10}$$

that must vanish. Hence, it follows that the mean surface temperature is:

$$T_s = \left[\frac{S_0(1-\alpha)}{2\sigma(2-\varepsilon_{\text{earth}})}\right]^{1/4} = \left(\frac{2}{2-\varepsilon_{\text{earth}}}\right)^{1/4} T_e \tag{5.11}$$

And the mean atmospheric temperature is:

$$T_s = \left(\frac{1}{2}\right)^{1/4} T_s \tag{5.12}$$

If the average emissivity, $\varepsilon = 0.8$, then $T_s = 290$ K and $T_a = 244$ K. This is close to the global mean surface temperature of 288 K.

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Chapter 6
Atmospheric Thermodynamics

6.1 Definition

Thermodynamics represents a branch of physics which deals with the energy and work of the system and the energy transfer from one place to another. Thus, thermodynamics is defined as the study of equilibrium states of a system which has been subjected to some energy transformation. The main task of atmospheric thermodynamics is to interpret the average properties of macroscopic properties of a system in equilibrium, and related to proportionally slow movements, utilizing mathematical-physical concept (Bohren and Albrecht 1998; Tsonis 2002; Zdunkowski and Bott 2004; Curry and Webster 2005; North and Erukhimova 2009; Hantel 2013; Khvorostyanov and Curry 2014). Atmospheric thermodynamics deals with the transformations of the energy in a system and between the system and its environment. The system is a typical example of the matter. The equilibrium state of the system can be fully determined by a small number of properties such as pressure, temperature, and volume. These properties are known as state variables, or thermodynamic variables. It can be applied in many scientific disciplines, including physics, chemistry, and biology. Thermodynamics plays an important role in our quantitative understanding of atmospheric processes in a wide range, from the smallest cloud microphysics process to the global atmospheric circulation (Iribarne and Godson 1973). The purpose of this chapter is to introduce the basic elements and relationships that exist in the thermodynamics. We first discuss about the ideal gas equation and its application in dry air, water vapour, and moist air. Then a brief description of a thermodynamic system is given. The following section concentrates on the relationship between the mechanical work performed by the system and the heat that gets in the system, as expressed in the first law of thermodynamics. Then several sections follow, pertaining to apply previously mentioned in the atmosphere. Finally, in this section, we focused on the second law of thermodynamics and the concept of entropy, laws that constitute the basis for derivation of some important relations in the field of atmospheric science.

V. Spiridonov, M. Curic, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_6

6.1.1 Thermodynamic System

In general, a system is a specific sample of matter. In the atmosphere, a parcel of air is a system. A thermodynamic system is a component or element of interest that we separate from the rest of the universe by "boundaries", called the environment. The sum of the system and environment is the universe. There are three types of system:

1. *Isolated*: a system that exchanges neither matter nor energy with the environment.
2. *Closed*: a system that does not exchange matter with the environment but which may exchange energy.
3. *Open*: a system that exchanges both matter and energy with its environment (see Fig. 6.1).

In an open system, mass and energy can be exchanged with its environment. A system is defined as closed when it exchanges energy but not matter with its environment and as isolated if it exchanges neither mass nor energy. In atmospheric thermodynamics, we assume that most systems are closed. It is justified when there is no interaction between the system and its environment.

This is approximately true when:

- The system is large enough to ignore mixing with its surroundings
- The system is part of large homogeneous system

A set of physical quantities that are used to describe the conditions of the system is referred to as a "state". Under the term "thermodynamic condition", we can specify the following variables, pressure, temperature, density, volume, entropy, etc., which we called the state variables.

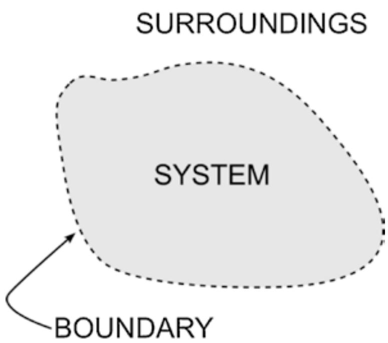

Fig. 6.1 Thermodynamic system

6.2 An Ideal Gas Law

Laboratory experiments show that pressure, volume, and temperature of a substance can be connected via the equation of state in a wide range of conditions. We require that all emissions approximately follow the same equation of state, which is called the equation of ideal gas. For many practical purposes, it can be assumed that atmospheric gases, whether they are considered as individual or as a mixture of gases, obey the following equation for ideal gas:

$$pV = nR^{*}T \tag{6.1}$$

where p is the pressure (Pa = kg m^{-1} s^{-2}), V is the volume (m^3), n is the chemical amount (in moles) which is equal to the total mass of the gas (m) (in grams) divided by the molar mass (M) (in grams per mole), R^* is the gas constant (8.314 J K^{-1} mole^{-1}), and T is the temperature in K, respectively. Usually, in the atmosphere, we do not know the exact volume of an air parcel or air mass. To solve this problem, we can rewrite the ideal gas law in a different useful form if we divide n by V and then multiply by the average mass per mole of air to get the mass density:

$$\rho = \frac{nM}{V} \tag{6.2}$$

where M is the molar mass (kg mole^{-1}). Density has SI units of kg m^{-3}. The value of R^* depends on the individual gas that is considered. Thus, from Eq. 6.1, the equation for ideal gas per unit mass (m = 1) is given by:

$$p = \rho RT = \rho \frac{R^{*}}{M} T \tag{6.3}$$

while $R = c_p - c_V$ is a specific gas constant of the observed ideal gas, which is related with the universal gas constant R^* through $R = R^*/M$, where M is the mass of one kilo mole of substance into SI system or one mole substance (molar mass).

If the density is expressed through the specific volume of gas $\rho = 1/\alpha$, then the following form of the ideal gas law is obtained:

$$p\alpha = RT \tag{6.4}$$

If the temperature is constant, Eq. 6.4 is reduced to Boyle's law, which says: if the temperature of a mass of gas is constant, the volume of gas is vice versa with pressure. Changes in the physical condition of the body that occur at a constant temperature are called isotherms. Also, implicit in Eq. 6.4, both Charles laws are given. The first of these laws is the constant mass of gas at constant pressure; the volume of gas is directly proportional to its absolute temperature. The second law is the constant

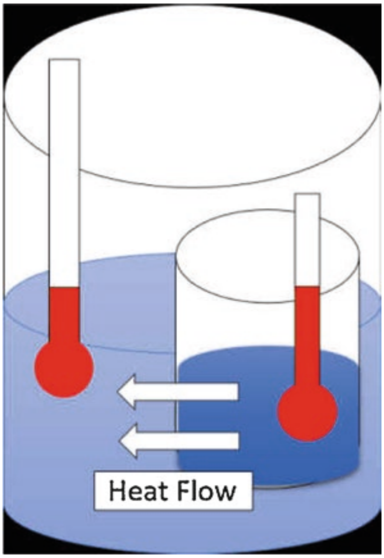

Fig. 6.2 Graphical illustration of the second law of thermodynamics. (Credot: BlyumJ [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

mass of gas contained in constant volume; the pressure of gas is proportional to its absolute temperature (Fig. 6.2).

6.2.1 The Equation of State of Dry Air

Often in meteorology we use mass-specific gas laws so that we must specify the gas that we are talking about, usually only dry air ($N_2 + O_2 + A_r + CO_2 + \ldots$) or water vapour (gaseous H_2O). We can divide R^* by M to get a mass-specific gas constant, such as $R_d = R^*/M_{\text{dry air}}$. Thus, we will use the following form of the ideal gas law for dry air:

$$p_d = \rho_d R_d T \tag{6.5}$$

where $R_d = R^*/M_{\text{dry air}} = 8.314$ J K^{-1} mole^{-1}/0.02897 kg mole^{-1} = 287 m^2 s^{-2} K^{-1} = 287 J kg^{-1} K^{-1}.

6.2.2 The Equation of State of Moist Air

The same procedure can be applied to water vapour:

$$p_v = e = \rho_v R_v T \tag{6.6}$$

where $R_d = R^*/M_{\text{water vapour}} =$ 8.314 J K^{-1} mole^{-1}/0.01802 kg mole^{-1} = 461 m^{2}s^{-2} K^{-1} = 461 J kg^{-1} K^{-1}. Typically, e is used to denote the water vapour pressure, which is also called the partial pressure of water vapour.

6.3 First Principle of Thermodynamics

Atmospheric thermodynamics deals with equilibrium states of the system. It basically states that the total energy of the system will remain constant if no energy is added or subtracted from the system. If we imagine a closed system in which the total amount of mass (either in the form of gas, liquid, or solid state or a mixture of phases) does not change, such thermodynamic system contains an internal energy due to kinetic and potential energy of its molecules or atoms. In this case, the system of unit mass has a certain amount of thermal energy (Q) which the system can obtain through thermal conductivity and/or radiation. Consequently, the system can carry out external work (W). The excess energy which the body obtains as the result of the external work is the difference between the thermal energy and the work that should be done by the system itself ($Q - W$). Following the principle of conservation of energy, the internal energy of the system must increase ($Q - W$), i.e.

$$Q - W = U_1 - U_2 \tag{6.7}$$

where U_1 and U_2 are internal energies of the system before and after the change. In differential form Eq. 6.7 becomes:

$$dQ = dW + dU \tag{6.8}$$

$$dQ = dW + (E_2 - E_1) = dW + dE \tag{6.9}$$

where dQ is the differential change of heat added to the system, dW is the differential change of the work carried out by the system, and dU represents the differential change in internal energy of the system. Equations 6.7 and 6.8 express the *first principle of thermodynamics,* whose graphical illustration is shown in Fig. 6.3.

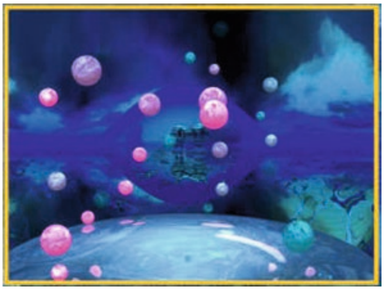

Fig. 6.3 Entropy of the system

Actually, Eq. 6.8 defines the change of internal energy of the system where dU depends only on the initial and final states of the system and does not depend on the way in which the system is transferred between these two states. Such parameters are known as *state functions*. The first principle of thermodynamics is an expression for maintaining the energy of the system. The first law of thermodynamics provides the basic definition of internal energy, associated with all thermodynamic systems, and states the rule of conservation of energy.

6.3.1 The First Principle of Thermodynamics for an Ideal Gas

For an ideal gas, internal energy is a function of temperature only, because the energy added by heating at constant volume only increases the accidental molecular movement, which is proportional to the temperature and is expressed as:

$$dU = C_V dT \tag{6.10}$$

After replacing $dW = pdV$ in Eq. 6.8, where $\alpha = \frac{1}{\rho}$ is the specific volume of gas, we obtain:

$$C_V dT = dQ - pd\alpha \tag{6.11}$$

Or by some arrangement:

$$dQ = C_V dT + pd\alpha \tag{6.12}$$

Thereby, we can use the equation of state for an ideal gas, to write the alternative form of the first principle of thermodynamics. The equation of state is the following:

$$pd\alpha + \alpha dp = RdT \tag{6.13}$$

Using the above relation, the first principle of thermodynamics (6.8) becomes:

$$dQ = \left(C_V + R\right)dT - \alpha dp \tag{6.14}$$

For isobaric process:

$$dQ = \left(C_V + R\right)dT \tag{6.15}$$

From the definition of heat capacity of air at constant pressure C_p, $dQ = C_p dT$ with replacement $C_V + R = C_p$, we get the alternative form of the first principle of thermodynamics:

$$dQ = C_p dT - \alpha dp \tag{6.16}$$

Or if $Q = H$, where H is a new quantity called enthalpy, then:

$$dH = c_p dT - \alpha dp \tag{6.17}$$

Enthalpy of the ideal gas as internal energy of gas is only a function of temperature. The other form is specific enthalpy:

$$dH = c_p dT \tag{6.18}$$

In meteorology, quantity $C_V dT$ is also called latent heat. It is the energy transferred to the heating system during isobaric process.

6.3.2 Enthalpy

The internal energy U and the volume V are not the only state variables that we can use to characterize a thermodynamic system. We can choose other quantities that can be closely related to U and V, such as the temperature T, pressure p, and enthalpy H. One commonly used state variable is called enthalpy and is defined as:

$$H \equiv U + pV \tag{6.19}$$

The differential of H is given as $dH = dU + pdV + Vdp$. This makes it possible to write the first law of thermodynamics as:

$$dH = dQ + Vdp \tag{6.20}$$

From the first form of the thermodynamics law, $dU = dQ - pdV$, we see that at constant volume, $dU = dQ$. Enthalpy of the ideal gas, as an internal energy of gas, is only a function of temperature. It is worth repeating the following:

1. For constant pressure processes, heat and enthalpy change are equivalent.
2. For constant volume processes, heat and internal energy change are equivalent.

6.3.3 Poisson Equations

Even though energy in the atmosphere/ocean system ultimately comes from radiative heating, adiabatic processes in the atmosphere are of interest for several reasons. Often it is because real atmospheric processes occur quickly in comparison with the time scale for heat transfer and so may be approximately adiabatic.

Alternatively, we may wish to make the adiabatic assumption simply because we are ignorant of the heat transfer and consequently must ignore it or give up. Poisson's equations describe relationships between the state variables T, p, and ρ for the adiabatic processes. When $dB = 0$ (adiabatic process), the first principle of thermodynamics can be written as:

$$c_v dT + p dV = 0 \tag{6.21}$$

Substituting for p from the ideal gas law, dividing by $C_V T$, and using $R = C_p - C_V$, we have, after some replacements:

$$d \ln\left(T V^{\gamma-1}\right) = 0 \tag{6.22}$$

where $\gamma = C_p/C_v$ (which is equal to 7/5 for an ideal gas). This implies that, for an ideal gas undergoing an adiabatic process:

$$T V^{\gamma-1} = \text{const.} \tag{6.23}$$

Equation 6.22 is one of the three Poisson's equations. Eq. (6.22) can be used to explain several atmospheric phenomena. Starting with the second version of the first law of thermodynamics, $dQ = C_V dT + p d\alpha$, and following a similar replacement, one may arrive at the second Poisson's equation for an adiabatic process in an ideal gas:

$$T p^{-k} = \text{const.} \tag{6.24}$$

where $k = R/C_p$ (which for an ideal gas is equal to 2/7). Finally, by combining Eqs. 6.22 and 6.23, we arrive at the third Poisson's equation

$$p V^{\gamma} = \text{const.} \tag{6.25}$$

where $\gamma = \dfrac{C_p}{C_V} = 7/5$. Equation 6.24 is the most familiar of Poisson's equations, although they are all equivalent.

6.3.4 Potential Temperature

Meteorologists use Eq. 6.24 to define a quantity known as potential temperature (a thermodynamic concept that seems to be unique to meteorology). Suppose we start with a parcel of air in some arbitrary initial state specified by T, p. Let us move the air parcel adiabatically to a pressure of 100 kPa and call the temperature which it achieves the potential temperature, θ. By using Eq. 6.24 in the initial and final states, it can be easily shown that:

$$\theta \equiv T\left(\frac{100}{p}\right)^{k} \tag{6.26}$$

where p must be expressed in kPa (since we have used 100 kPa in the numerator). Because the potential temperature of an air parcel is conserved under dry adiabatic processes, it may be used as a tracer for air parcels.

6.3.5 *Implementation of the First Principle of Thermodynamics*

We will calculate the specific energy added by heating (q), the specific work carried out (W), changes in internal energy (dU), and changes in enthalpy (dH), for isotherm, isobaric, and esoteric processes (constant volume) of an ideal gas.

Isotherm process: For $dT = 0$ at isothermal process, expressions of changes in specific internal energy of ideal gas ($dU = C_V dT$) and enthalpy ($dH = C_p dT$) are equal to zero. Since ($dU = 0$) for this process, the first principle of thermodynamics (6.8) simply becomes:

$$Q = W = \int_{\alpha_1}^{\alpha_2} p d\alpha = RT \ln\frac{\alpha_2}{\alpha_1} \tag{6.27}$$

The total energy is transformed in the system by heating during isothermal expansion of ideal gas, and it is used by the system to perform work in the area.

Isobaric process: In isobaric expansion of the gas, because the change in temperature $dT > 0$, changes in specific internal energy ($dU = C_V dT$) and enthalpy ($dH = C_p dT$) are also positive. The specific work carried out by the system is:

$$W = p\left(\alpha_2 - \alpha_1\right) \tag{6.28}$$

To calculate the energy delivered to the heating system, an alternative form of the first principle of thermodynamics is used:

$$dQ = c_p dT - \alpha dp \tag{6.29}$$

Since $dp = 0$ during isobaric process, the first principle of thermodynamics simply becomes $Q = dH$. The whole energy passing through the heating system during the isobaric process is used to increase the specific enthalpy of the system:

$$dQ = c_p dT - \alpha dp \Rightarrow c_p dT = dH \tag{6.30}$$

Isosteric process Since $dT > 0$ for isosteric process (constant volume), in which the pressure increases, changes in specific internal energy $dU = C_V dT$ and enthalpy $dH = c_p dT$ are also positive. The specific work carried out by the system is equal to zero, because ($d\alpha = 0$) during the isosteric process. The first principle of thermodynamics (6.8) simply becomes:

$$dQ = dU + pdV => dU \tag{6.31}$$

The overall amount of energy transferred through the heating system at isosteric process is used to increase the specific internal energy of the system.

6.4 The Second Principle of Thermodynamics

While the first principle of thermodynamics says that energy is conserved in all thermodynamic processes, the second principle of thermodynamics is concerned with the direction of natural processes and definition of a new variable to the state called "entropy". The second law claims that a natural process runs only in one sense and is not reversible. For example, heat always flows spontaneously from hotter to colder bodies, and never the reverse, unless external work is performed on the system (see Fig. 6.2). The explanation of the phenomena was given in terms of entropy. Total entropy (S) can never decrease over time for an isolated system because the entropy of an isolated system spontaneously evolves towards thermodynamic equilibrium: the entropy should stay the same or increase.

6.4.1 Definition of Entropy

Entropy has different physical interpretations, including statistical disorder (chaos) of the system (see Fig. 6.3) but for our purposes to consider the entropy as another feature of the system, as enthalpy or temperature. The second law says that there is a useful variable in the state called entropy (S). The change of entropy dS is equal to the transfer of heat dQ divided by temperature (T) which follows that the change of entropy is then the inverse of temperature:

$$dS = dQ / T \tag{6.32}$$

For a given physical process, the entropy of the system and the environment remains unchanged if the process is the return (reversible). If you mark the initial and final states of the system (i) and (f), then there is a balance between these situations:

$$S_f = S_i \left(\text{reversible process}\right) \tag{6.33}$$

Example of feedback process is ideally forcing the air flow through the narrowed tube. Ideally, that means no loss in the border layer. As the fluid flows and moves through the narrowing, the pressure, temperature, and speed are changing, but these variables returned to their original values downstream of the narrowing. The state of the gas returns to its original conditions and the change of entropy of the system is zero. This process is called isentropic process, i.e. the entropy of the system is constant. In order to calculate the change in entropy, we use the first principle of thermodynamics:

$$dU = dQ - dW \quad \text{or} \quad dQ = dU + pdV \tag{6.34}$$

Changing the definition of term which expresses the work of gas:

$$dQ = dU + dVc_p dT - \alpha dp \tag{6.35}$$

where

$$c_p dT = dH. \tag{6.36}$$

In the above expression, (p) is the pressure and (V) is the volume of gas. If we use the definition of the enthalpy of the gas (H):

$$dH = dU + pdV + Vdp \tag{6.37}$$

Enthalpy change for the next expression is obtained. By replacing in Eq. 6.8, an alternative way of presenting the first principle of thermodynamics is obtained:

$$dQ = dH - Vdp - pdV + pdV = dH - Vdp \tag{6.38}$$

For the ideal gas, the equation of state is written as $pV = RT$, where (R) is the gas constant. The transfer of heat of gas is equal to the heat capacity multiplied by the change of temperature, which in differential form is expressed as

$$dQ = C_V dT \tag{6.39}$$

If we consider the process of constant volume, the first principle of thermodynamics, the following expression is obtained:

$$dE = dQ = C_V dT \tag{6.40}$$

Similarly, for the process of constant pressure, the formulation of the first principle of thermodynamics gives:

$$dH = dQ = C_V dT \tag{6.41}$$

If we assume that the heat capacity is constant with temperature, we can use these two equations to define the change of enthalpy and internal energy. If we replace the value of pressure (p) in the equation of state of ideal gas and the definition of dE in the first equation of energy, we get:

$$dQ = c_V dT + RT\, dV / V \tag{6.42}$$

With the replacement value (V), the equation of state, and the definition of change of enthalpy (dH), the first principle of thermodynamics is recorded by the expression:

$$dQ = c_p dT - RT\, dp / p \tag{6.43}$$

By replacing dQ in the form of differential equation for entropy, we obtain:

$$dQ = \frac{C_V dT}{T} + R \;\; dV / V \tag{6.44}$$

and

$$dS = C_p dT / T - RT\, dp / p \tag{6.45}$$

These equations can be integrated by the condition "1" to that "2" as follows:

$$S_2 - S_1 = C_V \ln\left(\frac{T_2}{T_1}\right) + R \ln\left(\frac{V_2}{V_1}\right) \tag{6.46}$$

and

$$S_2 - S_1 = c_p \ln\left(\frac{T_2}{T_1}\right) + R \ln\left(\frac{p_2}{p_1}\right) \tag{6.47}$$

where C_V and C_p are the heat capacity at constant volume and pressure, and (ln) is a symbol for a logarithmic function. Depending on the type of process, the change of entropy of the gas may be determined. From Eq. 6.46, the process at constant volume, the second term in the equation is zero, because $v_2/v_1 = 1$. Then you can determine the value of specific heat at constant volume process. The first term of the equation can be considered as a contribution to the constant volume process and the second term as an additional shift caused by the change of volume.

6.4.2 Summary on Reversible and Irreversible Processes

As a summary, “reversible process” is the one performed in such way that the system and the local environment may be restored to their initial states, without producing any changes in the rest of the “universe”. A little thought should convince you that no real process is reversible. The abstraction of reversible processes provides a clean theoretical foundation for the description of real-world, *irreversible processes*. The logical question arises “What real-world processes” make things irreversible? Dissipative effects (e.g. viscosity, friction, inelasticity), i.e. processes for which the thermodynamic equilibrium for mechanical, thermal, or chemical equilibrium, are not satisfied. In atmospheric and ocean sciences, it is common to assume no dissipative effects (particularly for large-scale flows) and that motion of the atmosphere and oceans is isentropic. Viscous and frictional dissipation of course occur, but a good understanding of the dynamics can be obtained from inviscid theory.

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Chapter 7
Air Temperature

7.1 Air Temperature Definition

Air temperature is one of the most important elements of weather and climate (Saha 2008). By definition, temperature is a physical quantity which characterizes the degree of heating a physical body, which occurs as a result of accidental secondary movement of molecules in the body. In a physical sense, temperature is a measure of the mean velocity of the level of kinetic energy of molecules. In other terms air temperature is characterized by the relationship of two bodies that are in mutual thermal contact tending to reach the same temperatures. More significant, air temperature is a measure of the atmospheric heat content as a response to combined effects of absorbed solar radiation by the Earth's surfaces, the vertical fluxes of sensible and latent heat released to the air by the process of convection, and horizontal advection (movement) of warm and cold air masses (Ambaum 2010; Lutgens and Tarbuck 2009). Thus, the temperature is a thermodynamic condition of the body, and its value is determined by the total flow of heat between bodies. In such systems, the overall body loses heat in relation to another body in case of a higher temperature. To define temperature as a physical quantity which is related to the state of the body is hard. For this purpose, an internationally accepted temperature scale is defined, based on universal freezing and the triple point of water. The temperature of the atmosphere is controlled by a complex system of interactions between biosphere, atmosphere, and lithosphere (Fig. 7.1). Energy is continuously exchanged between the surface and the air over the place and the circulation of air around the globe.

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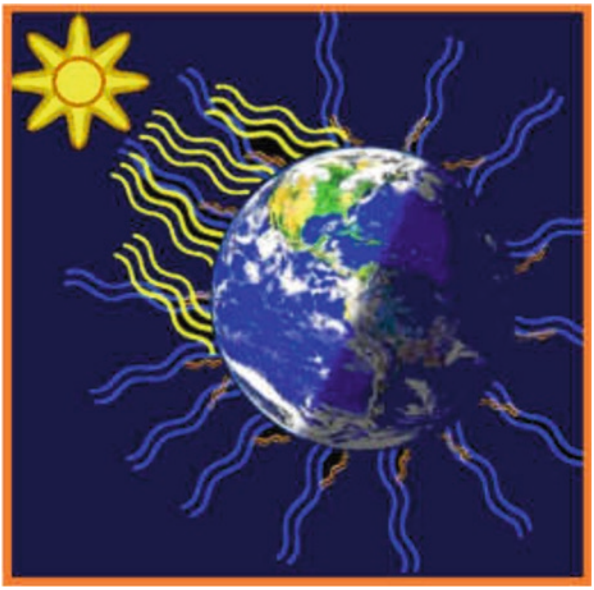

Fig. 7.1 Heat transfer in the atmosphere

7.2 Heat and Temperature

Heat is a measure of energy of matter. Matter consists of atoms and molecules which are constantly oscillating. Heat is the total kinetic energy of atoms and molecules. Unlike heat, temperature is a measure of the intensity or the mean kinetic energy of individual atoms and molecules. Heat and temperature are variables that are closely related to each other. By adding heat, the molecules are moving faster, and this contributes to increasing temperature. Conversely, by subtracting the heat, the molecules move more slowly, i.e. we say that the temperature decreases. The quantity of heat depends on the mass of matter (as total energy), but on the other hand, the temperature does not depend on it.

7.2.1 Heat Transfer

The second known principle of thermodynamics claims that all systems tend to create some disorder. Where the temperature gradient exists (change of temperature with a given distance), heat will flow in the direction of reducing the gradient (and of course speed will increase with the gradient). Therefore, heat is always transferred from the body with higher temperature to the body with lower temperature. There are three primary mechanisms of heat transfer: conductivity, convection, and radiation (Lutgens et al. 2018).

Table 7.1 Thermal conductivity of certain media

Material	(cal/sec)/ (cm^2 C/cm)	(W/mK)
Copper	0.99	385.0
Water (20 °C)	0.0014	0.6
Air at (0 °C)	0.000057	0.024

- Conductivity
- Convection
- Radiation

Conductivity Conductivity means transfer of thermal energy in unit time and per unit area of matter per unit temperature gradient (temperature difference). The ability to transfer heat through conductivity changes dramatically between substances and is expressed as the transfer of heat through the middle of a unit temperature gradient. For example, different media have different thermal conductivity. For example, water has greater thermal conductivity than air (Table 7.1).

The air is very low conductor of heat. Heat conductivity is measured in (cal/sec)/(cm^2 C/cm) or (W/m K).

Convection Convection is by definition the transfer of heat by movement within the fluid, whether a liquid or gas. When the fluid is heated, it expands and becomes less dense and more buoyant, and as such is lifting. At the same time, cooler fluid is denser and starts sinking. Heated air rises and cold air descends. Convection is a very important process that occurs in the lower part of the atmosphere. It plays a crucial role in small-scale currents, the formation of thermals, which create turbulent movements in the air.

Radiation Radiation means heat transfer which does need a medium. Radiation is a form of energy transport consisting of electromagnetic waves traveling at the speed of light. Radiation is the mechanism in which heat is transferred through space in a vacuum from the Sun to the Earth. Radiation is the transfer of electromagnetic energy through an electric or magnetic wave. When this energy is absorbed by the object, it appears to increase the molecular movement and temperature.

7.3 Temperature Factors

The primary factors (controllers) that affect the distribution of heat and therefore energy and cause the air temperature change from place to place (Ahrens and Henson 2016; Spiridonov and Ćurić 2010) are the following:

- Radiative transfer (different heating land and water)
- The movement of air masses

Fig. 7.2 Temperature factors (controllers)

- Geographical location
- Ocean currents
- Cloud cover
- Height above sea
- Albedo

A graphic illustration of air temperature factors (controllers) is shown in Fig. 7.2.

Solar radiation is the main source of heating of the Earth's surface. On the other hand, longwave Earth radiation is heating the air. While daily insolation and net radiation show similar daily patterns, they differ in magnitude. Air temperatures respond by generally increasing, while net radiation is positive and decreasing when net radiation is negative. Daily insolation tends to peak at noon, as does net radiation. Net radiation is positive for the entire 24 h period in the summer, while it is negative in the winter. Daily insolation is much greater in summer because the Sun is higher in the sky and the daylight period is much longer than in the winter.

The movement of air masses and geographical location are very important factors that affect the air temperature. Temperature decreases from the equator polewards. Isotherms suffer latitudinal departure with the seasons. The highest and lowest air temperatures appear on land.

Clouds are good absorbers of Earth terrestrial radiation. Air temperature depends on the coverage of the atmosphere with clouds and presence of emissions. Greater cloud cover (cloudiness) means less incoming solar radiation and consequently reduces the air temperature.

Land cover elevation is heated faster than water. Great heat capacity of water (oceans, seas, and lakes), which constitute about 2/3 of the Earth, provides them to accumulate greater quantities of heat and affects the heat balance of Earth. The surface of the land versus water absorbs less, but again, faster and re-emitted heat. The water absorbs more heat and pays more slowly, so that land quickly loses heat and cools quickly at night. Water loses heat very slowly, so that in areas where large water bodies are present, there are no sudden changes in temperature. The other key factor in determining surface temperature is elevation. Surface temperature declines ~1 °C for every 220 m in elevation above sea level. The coldest portions of Earth are the Greenland and Antarctic Ice Sheets, which combine both very high latitude and high elevation.

Geographical location is also an important factor that affects temperature. Temperature decreases from the equator to the poles. Isotherms suffer latitudinal deviation with the seasons. The highest and lowest temperatures of the air can be noticed on the mainland. With increasing height above the sea level, the temperature decreases. The vertical temperature gradient is greater than the horizontal temperature gradient.

Albedo is important temperature controller in the degree of reflection of short-wave solar radiation that reaches the Earth's surface. Earth's albedo is one of the elements that determine the surface temperature, but also determine the climate. Albedo is a function of different elements that make up Earth's surface and atmosphere: clouds, deserts, glaciers, oceans, forests, and others. For example, forests or oceans reflect 5 to 15% of received radiation, desert sand 30%, and glaciers 50%, but the fresh snow reflects up to 98% of solar radiation. By increasing height above the sea, temperature is decreasing. The vertical temperature gradient is larger than the horizontal temperature gradient.

7.3.1 Heat Advection

According to standard definition, thermal (or heat) advection is the transport of latent heat by a moving fluid (bulk motion), such as air. When the wind blows from a warmer to a colder region, there is warm advection. The heat advection depends on wind and the temperature gradient. Figure 7.3 shows an example of heat advection illustrated through the clouds, which is manifested through the penetration of cold air (blue colour) in the region with higher air temperature (red). A state of the atmosphere in which density depends only upon pressure, that is, a state such that surfaces of constant pressure and constant density coincide, so that the geostrophic wind is independent of height.

The advection occurs as the result of some external forcing such as wind that causes horizontal air particle displacement. Depending on the air mass distribution and the atmospheric systems movement, the warm or cold advection occurs. For example, a warm frontal system is associated with a warm heat advection over the surface towards a cold air mass. Oppositely, cold front moves a cold air mass into a

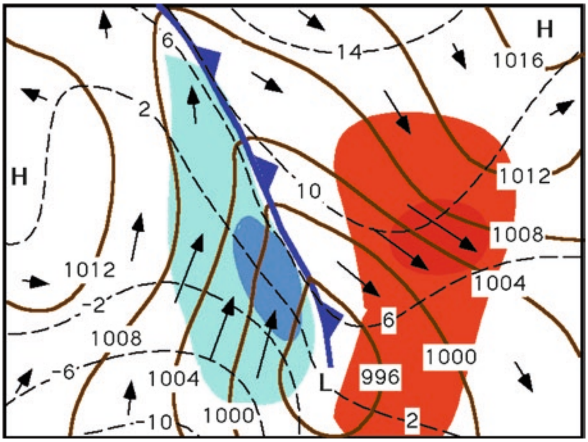

Fig. 7.3 Cold and warm advection
(a) The thick solid lines are 42 hPa. The geostrophic wind (which is shown by arrows) at the Southern Hemisphere blows along these contours. The length of the arrows is a measure of the wind speed. The dashed lines are isotherms (°C). Blue shading indicates cold advection; red shading is warm advection. (b) A cold front extends from a low towards the north. The density of the shading is proportional to the amount of advection, i.e. the darkest area experiences the strongest advection (Source: http://www-das.uwyo.edu/~geerts/cwx/notes/chap03/advection.html)

warmer region. There is also heat advection into the ocean water as means of energy exchange between atmosphere and ocean due to the external forcing as radiation or wind-driven vortex currents.

7.4 Temperature Changes

Rotation of the Earth is a major factor for the daily changes of temperature. The magnitude of these changes depends on local factors, i.e. local weather conditions. Because of heating the Earth's atmosphere, the months with highest and lowest temperatures here do not coincide with periods of maximum and minimum incoming solar radiation. Although the largest intensity of solar radiation occurs during the summer solstice, the Northern Hemisphere is usually the warmest in July and August. Conversely, the Northern Hemisphere, with the minimum solar radiation arriving in December during the winter solstice, January and February are usually colder.

7.4.1 Daily Insolation and Pure Radiation

Although daily insolation and net radiation showed similar daily distribution, they differ in amount. Insolation is strictly determined by net radiation. Air temperatures increase when net radiation is positive and decrease when it is negative. In both cases, in high latitudes, insolation and net radiation are related to season and length of day, as opposed to night-time. Daily insolation aims to reach a maximum at noon, as the net radiation (Fig. 7.4). Net radiation is positive for all 24 h period in the summer, while it is negative in winter. Daily insolation is much higher in summer because the Sun is high in the sky and the day is much longer than in winter.

7.4.2 Daily Air Temperature Distribution

Incoming solar radiation significantly changes as a result of the rotation of the Earth. During the day, the total radiation is positive, and the Earth's surface receives heat. During the night radiation is negative, and the surface loses heat through the Earth's longwave radiation in the atmosphere and space. This results in the appearance of the daily cycle of increasing and decreasing the temperature in the air (Ćurić and Janc 2016). Daily minimum air temperature usually occurs about half an hour after the sunrise. When pure radiation becomes positive, the surface is quickly heated and transferred to heat in the air that lies above it.

Air temperature increases sharply in the morning and extends the same pace after the midday maximum solar radiation. This natural temperature interval, to a certain extent, appears to violate the convection (transport of warm air in the upper atmosphere), in the afternoon in the warm part of the year. When the Sun sets, the temperature decreases rapidly and continues to decline with some reduced rate through the night. Daily mean air temperature is determined by averaging the 24 h measurements (readings) or by collecting the maximum and minimum temperature for the 24 h averaged period. Daily temperature range is calculated by finding the difference between maximum and minimum air temperature. Data from the daily hour measurements of temperature are saved (stored) and are used for further processing. Depending on the nature of the temperature and the statistical treatment period (month or year), the following established definitions of temperature is applied:

(a) Monthly mean temperature (sum of the average daily temperatures for each day of the month, divided by the number of days in the month)
(b) Annual mean temperature (the average of 12 monthly mean temperatures)
(c) The annual air temperature amplitude (the difference between the warmest and coldest monthly mean temperature)

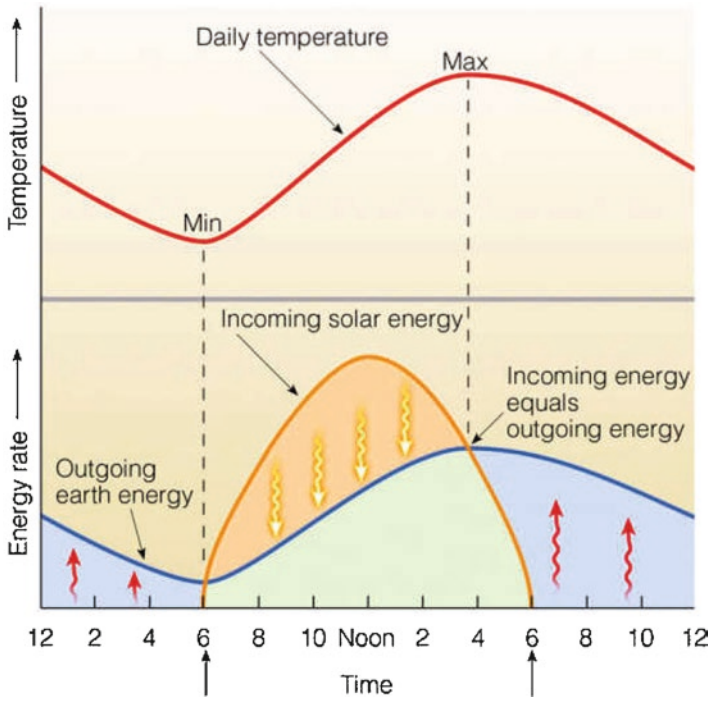

Fig. 7.4 Daily cycle of solar radiation and temperature

7.4.3 Vertical Temperature Change

In the standard atmospheric conditions, the air temperature decreases with height in the lower layer of the atmosphere. Figure 7.5 shows the vertical profile of temperature in standard atmosphere. The troposphere is the layer of the atmosphere closest to the Earth's surface where the weather takes place. The warmest temperatures in the troposphere are near the surface with the coldest temperatures being at the top of the troposphere. Although the sunlight comes from the top to the bottom of the atmosphere, the troposphere is primarily heated from the bottom. This is because the surface is much better at absorbing a wide range of solar radiation as compared to the air. The surface is warmed by the Sun, and then this energy is distributed upwards into the troposphere through a mixing of the air.

Since the Earth's surface is the primary heat source, temperatures will be warmest at the surface and decrease away from the surface. The average temperature profile of the troposphere will show a decrease in temperature with height. A vertical temperature gradient is a physical quantity that describes the rate at which temperature changes the most rapidly in vertical direction. The temperature gradients in

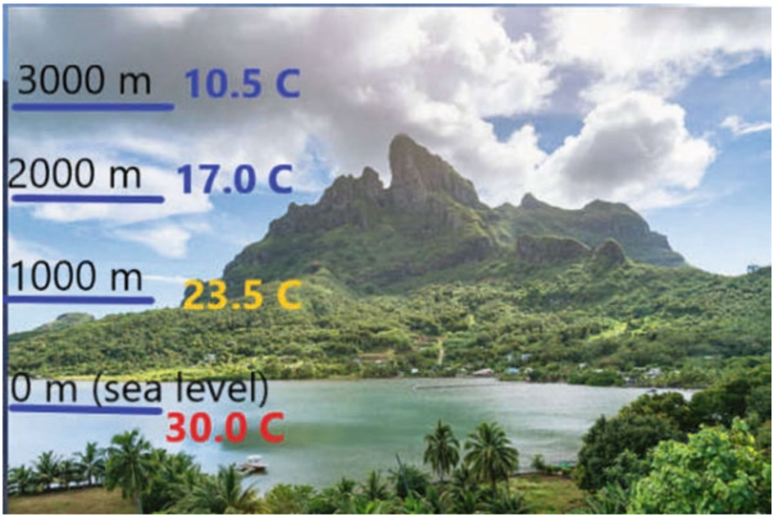

Fig. 7.5 Vertical temperature profile

the troposphere change with the geographic latitude, the climate zone, and season. The exact rate at which the temperature decreases with height – the environmental lapse rate varies with location and daily conditions. The changes are generally smaller than the interval (0 °C/100 m) during winter than half, reaching more than −0.8 °C/100 m over the summer subtropical ocean. In the middle latitudes, the vertical gradient of temperature has different amounts, but its average is decreasing about −0.65 °C/100 m. In the troposphere temperature decreases with height due to radiation cooling. On the other hand, the surface is heated as a result of solar radiation. Temperature changes slowly with altitude in the clouds, because of the process of condensation and the release of latent heat. In the stratosphere, temperature increases by about −0.2 °C/100 m with an increasing altitude. This occurs due to radiation heating which occurs because of the reaction of ozone with ultraviolet radiation, whereby heat is released. Under certain atmospheric conditions inside each tropospheric layer, air temperature may rise with an increasing altitude. This reverse condition is called temperature inversion. The vertical temperature gradient in this case is negative. In the mesosphere again exists the process of radiation cooling where the temperature is reduced by about −0.2 °C/100 m. When the air temperature in the layer remains constant with height, then there is an isothermal layer. During the night, especially when the sky is clear, the intervention significantly cools down in the air layer, but longwave radiation from higher layers is weak, and the air cools down to only about 1 °C. Consequently, a nocturnal inversion layer above the mixing is established.

7.4.4 Adiabatic Changes of Temperature

As air is heated, it expands becoming less dense and starts rising upwards above the cooler air. This process occurs without exchange of heat with the surrounding air (Barry and Chorley 2003). The lapse rate at which air temperature changes is called adiabatic change of temperature. The rate of cooling or heating of unsaturated "dry" air, moving vertically upwards, is (−1.0 °C/100 m) and is called "dry adiabatic gradient". At the condensation level of air rising (height at which the air achieves saturation and the formation of clouds begins), latent heat is released, and the rate of cooling is reduced. Slower rate of cooling, known as wet adiabatic gradient ("wet" because the air is saturated), changes from 0.5 °C/100 m of air with high moisture content up to 0.9 °C/100 m for air with low moisture content.

7.4.5 Temperature Inversion

Under normal conditions, in the troposphere in the near surface layer, the air temperature decreases with increasing height. Conversely, the temperature inversion is a thin layer of the atmosphere, where the temperature rises with altitude. Inversion, also called "stable" atmospheric layer, occurs under stable atmospheric conditions and the negative vertical temperature gradient. Inversions play an important role in determining cloud formation, precipitation, and visibility. An inversion acts as a cap on the upward movement of air from the layers below. Inversions also affect diurnal variations in air temperature. The principal heating of air during the day is produced by its contact with a land surface that has been heated by the Sun's radiation. The graphic illustration of the phenomenon of temperature inversion in winter is shown in Fig. 7.6. In terms of temperature inversion during winter in valleys, the appearance of fog is usually typical. Above the level of inversion, the weather is sunny, shining, and relatively warmer, with luminous atmosphere.

7.4.6 Global Distributions of Temperature

Analysis of global maps shows decrease in the average air temperature at sea level, from the equator towards the poles. Maximum temperature is evidenced on the south of the equator in January and July in the north. The coldest and warmest regions are those over land. Minor change in temperature distribution along the geographic longitude exists in the Southern Hemisphere. Isotherms also affect the ocean currents. The difference between the 2 months shows less seasonal change in temperature near the equator. Farther from the tropics, there is a more pronounced seasonal change of air temperature on the Earth's surface than above the ocean.

Fig. 7.6 Thermal inversion during winter

7.4.7 Surface Temperatures

The near surface temperatures are on average more extreme than air temperatures at a large distance from the surface. There is an evident difference between urban and rural temperatures. In rural areas, transpiration of plants removes large quantities of heat from the surface. Evapotranspiration of water from moist soil surfaces also cools the surface much more than in the average city. Urban areas are covered with a different infrastructure, buildings, and other objects. Much less vegetation is present, so transpiration occurs at a much lower rate. City surfaces are also darker and more absorbent than rural surfaces. Heat absorption is enhanced by the many vertical surfaces in cities, which reflect radiation from one surface to another. Concrete, stone, and asphalt conduct and hold heat better than soil, even when the soil is dry. Fuel consumption and waste heat also contribute to increased temperatures.

7.4.8 Urban Heat Island

As a result of these effects, air temperatures in the central region of a city are typically several degrees warmer than those of the surrounding, countryside, and rural areas. This has important economic and environmental consequences; higher temperatures demand more air conditioning and power consumption, and higher temperatures enhance smog formation.

7.5 Temperature Scales

Temperature scales are the basic units for the quantitative extent (determining the values) of temperature. Temperature scales use different reference points, sometimes known as fixed points. There are three main temperature scales used (Fig. 7.7):

1. Fahrenheit scale, which is determined using the point of ice (32°) and the water vapour point (212°)
2. Celsius scale, decimal scale at which the melting point of ice is set to 0 °C and boiling point of water at 100 °C

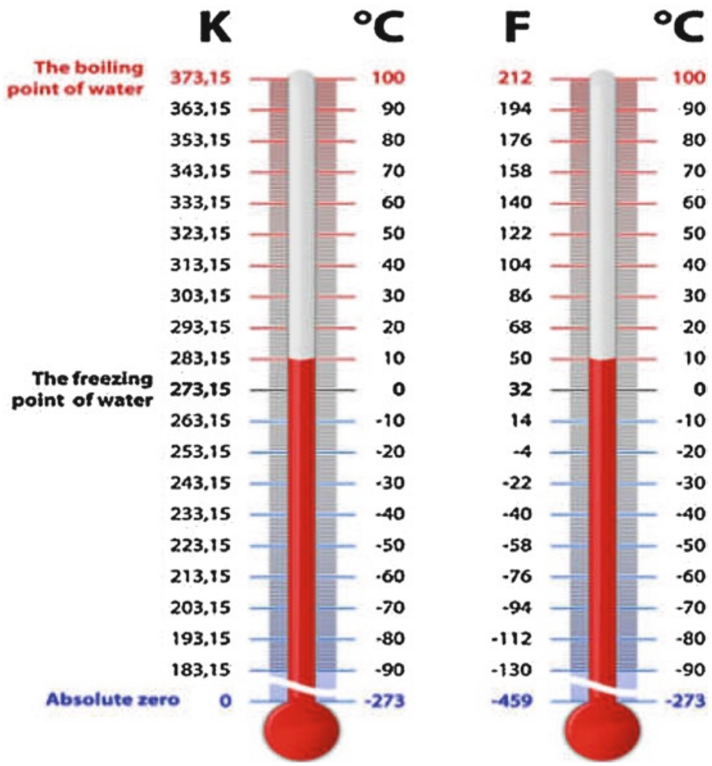

Fig. 7.7 Basic temperature scales

3. Kelvin scale (is an absolute temperature scale). Absolute zero, or 0 °K, is the temperature at which molecular motion is a minimum. Absolute zero corresponds to a temperature of −273.15° on the Celsius temperature scale. The ice point in the Kelvin scale is set at 273 K, while the boiling point is the 373 K

7.6 Isotherms

Isotherms are lines connecting points with the same value of temperature. For example, drawn isotherms on the maps illustrate the distribution of the temperature of the globe, with a general line's direction, east and west. Furthermore, isotherms illustrate the displacement of temperatures in latitude, as a result of seasonal movement of the vertical rays of the Sun. The global averaged isotherm distribution for January and July clearly shows the displacement of temperatures in latitude, as a result of seasonal movement of the Sun radiation. They also indicate the presence of ocean currents. The displacement of isotherms from north to south is more pronounced over continents, because the temperature deviation is higher over continent than over the water.

7.7 Latitudinal Heat Balance

The tropics have the most direct sunlight and highest temperatures. While the seasonal contrasts in surface temperature are due to the tilt of the Earth axis, there is relatively little variation in the annual average sunlight received throughout these entire tropics, and hence the entire band has similar temperatures. The other key factor in determining surface temperature is elevation. Surface temperature declines ~1 °C for every 220 m in elevation above sea level. The coldest portions of Earth are the Greenland and Antarctic Ice Sheets, which combine both very high latitude and high elevation. The shift to vertical rays means that in an average year, tropical latitudes receive more solar insolation than polar latitudes. At the same time, the Earth's radiation changes with the latitude due to temperature (T) variations. Specific latitudes have no global balance between incoming and outgoing radiation. Incoming radiation is higher than the output radiation at lower latitudes (see Fig. 7.8), while the outgoing radiation is greater than input at higher latitudes. This should mean that tropics continue heating until the pole is cooling. But that does not happen because the atmosphere and the oceans are machines that transfer heat from the tropics to the poles. Alternatively, latitudinal heat balance raises the atmospheric and the oceanic circulation or manages them. Coldest and warmest regions are those over the land. There are minor changes in geographical latitude in the Southern Hemisphere. Isotherms affect the ocean's currents. Also, the difference between the 2 months shows less seasonal changes near the equator and outside the tropics and more seasonal changes over the land than over the ocean.

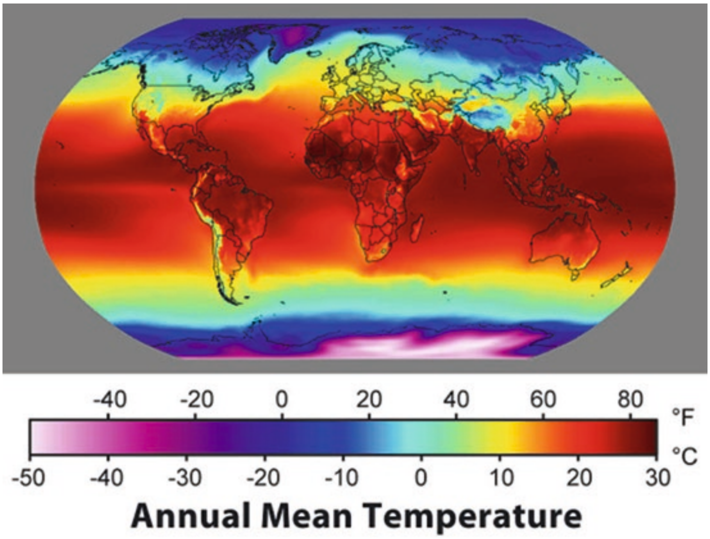

Fig. 7.8 A global map of the annually averaged near surface air temperature from 1961 to 1990. (Credit: Robert A. Rohde https://en.wikipedia.org/wiki/User:Dragons_flight)

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Chapter 8
Atmospheric Pressure and Wind

Atmospheric pressure is the weight of the air column that lies above the unit surface. This pressure is only partly influenced by the air velocity of the column. The value of air pressure changes both in space and in time. Pressure changes are greatest in the vertical direction, and close to sea level, they are about 1 mb every 10 m. In the horizontal direction, the typical pressure change is about 1–2 mb per 100 km. Despite the horizontal pressure changes are far less than vertical (at moderate latitudes about ten thousand times), they are of great importance for the dynamics of the atmosphere (e.g. Ackerman and Knox 2007; Markowski and Richardson 2010; Ćurić and Janc 2016; Houghton 2002; Lindzen 2005; Lutgens et al. 2018; Sanchez-Lavega 2011; Spiridonov and Ćurić 2011). Changes in atmospheric pressure in the horizontal direction and time can occur for both thermal and dynamic reasons. The cooling of the air causes the atmospheric pressure to rise, because the cooler air is denser and therefore heavier, while the value of the ground pressure decreases as the air warms, as it becomes less frequent and lighter. The air in the atmosphere is in constant motion. The penetration of cold or warm air causes changes in the atmospheric pressure in the observed area. In areas with upward air motion, the surface pressure decreases, while in areas downstream pressure increases. Atmospheric pressure is an indicator of weather. When a low-pressure system moves into an area, it usually leads to cloudiness, wind, and precipitation. High-pressure systems usually lead to fair, calm weather. If a barometre shows low barometric pressure tendency and shows air pressure tendency increase, it suggests that a high-pressure system is taking over and fair weather conditions are expected.

8.1 Mass of the Atmosphere

One of the possible ways to determine the mass of the atmosphere (m_a) is to utilize the basic physical definition of air pressure as force (F), which acts on unit area (S) on the Earth's surface:

V. Spiridonov, M. Ćurić, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_8

$$p = \frac{F}{S} \tag{8.1}$$

and the expression of the weight of the surrounding atmosphere, which is:

$$G = m_a g \tag{8.2}$$

In the case when (F) is equal to the weight of the atmosphere (G), it follows that:

$$p_{sfc} S = m_a g \tag{8.3}$$

Pressure (p), mass (m), and the acceleration due to gravity (g) are related by $p = \frac{F}{S} = mg / S$. Atmospheric pressure is thus proportional to the weight per unit area of the atmospheric mass above that location. Here, if the Earth's surface is replaced ($S = 4\pi R^2$), where ($R = 6378$ km) is the radius of the Earth and the average pressure on the Earth's surface is $p_{sfc} = 984$ hPa, the following expression is obtained for the mass of the atmosphere:

$$m_a = \frac{4\pi R^2 p_{sfc}}{g} = 5.129 \times 10^{18}\,\text{kg} \tag{8.4}$$

8.2 Definition of Atmospheric Pressure

The air pressure is the force per unit area exerted on a surface by the weight of air above it. Or more simplified definition, the air pressure is the weight exerted by the overhead atmosphere on a unit area of surface. Air pressure that is exerted by the atmosphere found in rest indicates the weight of column of air that lies above it. It is usually called as "barometric pressure", which refers to the pressure within the atmosphere of Earth (see Fig. 8.1).

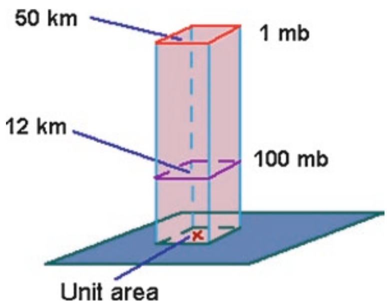


Fig. 8.1 An example of atmospheric pressure

Therefore, pressure decreases with altitude as there is less overlying weight of air. There are three ways in which pressure decreases with height:

1. Air rising
2. Decreasing the air density
3. Decreasing the air mass (i.e. upper level divergence)

The standard atmosphere (symbol, atm) is a unit of pressure mainly used as a reference value for the average atmospheric pressure at sea level. On average, the atmosphere of each (m^2) of the sea surface presses with an average force of about 10 tons or 10,000 kg/m2, while the total pressure of the atmosphere is about 100,000 N/m^2 or the average pressure at mean sea level (MSL) in the International Standard Atmosphere (ISA) is 1013.25 hPa, or 1 atmosphere (atm), or equivalent to 760.001 mmHg, at 0 °C. In most conditions atmospheric pressure is closely approximated by the hydrostatic pressure, caused by the weight of air above the measurement point. As elevation increases, there is less overlying atmospheric mass, so that atmospheric pressure decreases with increasing elevation. Atmospheric pressure is caused by the gravitational attraction of the planet on the atmospheric gases above the surface and is a function of the mass of the planet, the radius of the surface, and the amount and composition of the gases and their vertical distribution in the atmosphere. It is modified by the planetary rotation and local effects such as wind velocity, density variations due to temperature, and variations in composition. In meteorology atmospheric pressure is expressed in millibars or Pascals. Surface pressure is approximately 1000 mb (1013.25 mb); the total pressure is expressed through Dalton's law as a sum of partial pressures made by individual gases. The pressure is exerted equally in all directions, not just downwards, and always decreases with height. Additionally, the pressure is a function of density and temperature of the air. Atmospheric movements appear to equalize pressure differences.

8.3 Geopotential

The atmosphere is in the force field of Earth's gravity. Therefore each individual air particle in the atmosphere has potential energy. In order to raise some of the air to some height, work must be done, which in nature is usually done by the atmosphere, reducing its own amount of energy. The potential energy of air mass *m* at elevation *h* relative to the mean sea level is called *geopotential* Φ that is given with the following relation:

$$\Phi(h) = \int_0^h g(\phi, z)\, dz \tag{8.5}$$

where $g(\phi, z)$ is acceleration due to gravity, ϕ latitude, and z geometric height. It represents the potential energy of a unit mass relative to the mean sea level, and its unit in the **SI** system is m^2s^{-2} or a potential of Earth's gravity. It is a scalar quantity independent of air mass, which depends primarily on the height and latitude of the air parcel. Geopotential height is:

$$Z_g(h) = \frac{\Phi(h) - \Phi(0)}{g_0} \tag{8.6}$$

where g_0 is standard gravity at mean sea level. Geopotential increases with height, and in each direction where gz is constant, its value does not change. The surface where it is:

$$\Phi = gz = \text{const} \tag{8.7}$$

is called the equipotential surface. If the geopotential changes by 1 m^2s^{-2} ($\Delta\Phi = 1$), using the standard value of gravitational acceleration, the change of height will be:

$$\Delta Z = \frac{1}{g_0} \approx 1\text{dm}$$

This fact imposes the need to introduce a new unit for the height of the equipotential surfaces. Since these are horizontal surfaces, the gravity of Earth does not perform the work. Under these conditions, a unit of height is introduced in meteorology, called a geopotential meter (gpm). The geopotential height is connected to meters via a relation $\Delta Z = \frac{g}{g_0}\Delta z$, where ΔZ and Δz are heights expressed in geopotential metres and geometric (dynamic) metres, respectively. When $g > g_0$, the geopotential metre ($\Delta Z = 1$ gpm) is larger than the ordinary metre and vice versa, while they are the same for the standard value of the acceleration of the Earth's gravity. Thus, in horizontal air movement, the height expressed in (gpm) will not change following the equipotential surface, while those expressed in unique metres will change, except in the west-east direction. This fact is illustrated in Fig. 8.2. Values of geometric height, height in geopotential metres, and acceleration due to Earth's gravity (g) at 40° are shown in Table 8.1.

8.4 Barometric Pressure Distribution

In the weather analysis and forecast, it is very important to display the pressure field for the selected period (synoptic term) at both mid-sea and at upper levels. At the mid-sea level, the pressure field is represented by the so-called isobaric charts, while at altitude it is plotted by isohypses of a certain isobaric surface.

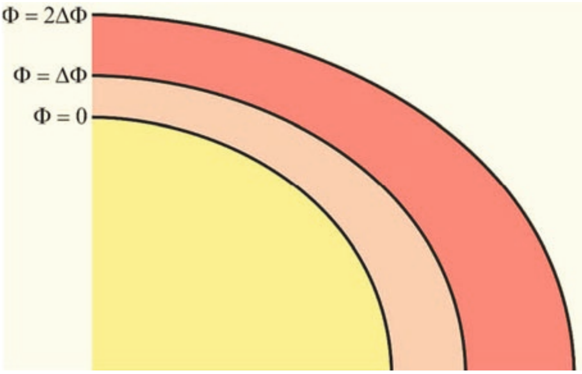


Fig. 8.2 Mutual positions of equipotential lines in the meridional plane

Table 8.1 Values of geometric height (z), height in Geopotential metre (Z), and the acceleration due to the gravity of Earth denoted (g) at 40° latitude

z (km)	Z (km)	$g = \text{ms}^{-2}$
0	0	9802
1	1000	9798
10	9986	9771
20	19,941	9741
30	29,864	9710
60	59,449	9620
90	88,758	9531
120	117,795	9443
160	156,096	9327
200	193,928	9214
300	286,620	8940
400	376,370	8677
500	463,597	8427

The maps show areas of high- and low-pressure gradients as well as different types of pressure systems. Steep pressure gradients indicate closely distributed isobars, while widely deployed isobar is an indication of weak gradients. The speed of the air flow is proportional to pressure gradients.

8.4.1 Pressure Gradient

The value of air pressure changes both in space and in time. Pressure changes are greatest in the vertical direction, and close to sea level, they amount to about 1 mb every 10 m. In the horizontal direction, the typical pressure change is about 1–2 mb per 100 km. Despite the horizontal pressure changes are far less than vertical (at moderate latitudes about ten thousand times), they are of great importance for the dynamics of the atmosphere. Changes in atmospheric pressure in the horizontal direction and time can occur for both thermal and dynamic reasons. The cooling of the air causes an increase in surface atmospheric pressure, as the cooler air is denser and therefore heavier, while the value of the surface pressure decreases as the air is heated, as it becomes less dense and lighter. The air in the atmosphere is in constant motion. The penetration of cold or warm air causes changes in the atmospheric pressure in the observed area. In areas where upward air movements exist, the pressure on the ground floor decreases, and in the downstream areas, the pressure increases.

8.4.2 Isobars

In meteorological practice, especially for the weather forecast, usually a field of pressure over a specific area of horizontal surface can be displayed using the isobar. Isobar lines are connecting points of equal atmospheric pressure or lines that join, on a weather map, areas of equal barometric pressure (Fig. 8.3). These lines are drawn on the map at intervals of 4 hPa (mb), above and below the value of 1000 hPa (mb). When the isobars are drawn in, they form definite patterns. They never cross but form roughly concentric circles and form themselves into distinct areas of high and low pressure. They have the same meaning as the height of the contours which show the geographical map. In synoptic meteorology isobars usually draw in different intervals (2, 4, or 5 hPa) depending on the size of the analysed area, the type of chart (surface or upper level), and the need for closer synoptic display field pressure. By definition, isobars cannot intersect each other. The map from which isobars are

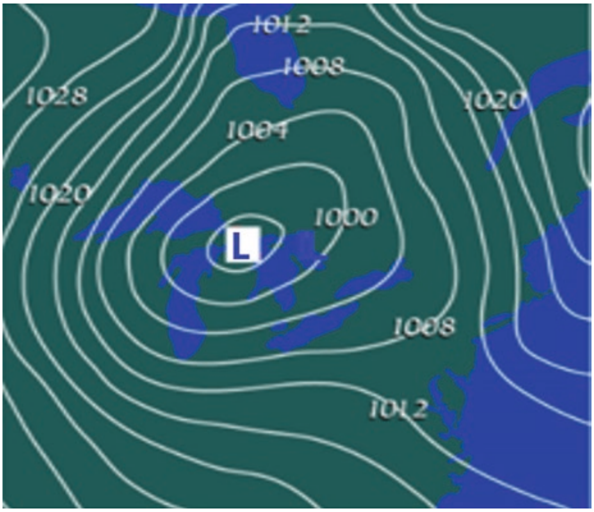


Fig. 8.3 Distribution of the field of atmospheric pressure at ground level that shows the isobar interval is 4 mb

drawn is called isobaric chart. Isobaric area is the area with a uniformed or a constant value of pressure.

Drawn isobars usually have some standard patterns. A system of curved isobars, covering an area of low-pressure depression, shows depression or cyclone. Isobars can also be considered as lines where individual isobar surfaces (points in space where the pressure is the same) intersect the horizontal surface (in this case the mean sea level), as shown in Fig. 8.4. From the figure it can be seen that the cyclone, i.e. the area of closed isobars in which pressure decreases from the periphery towards the center, in fact represents a valley which indicates low-pressure area, while areas with high pressure represents a high-pressure ridge, as it is shown in a topographic map. When isobaric fields are filled with information collected from meteorological stations in the form of weather symbols, forecasters can use their skill in order to provide a weather forecast over the next few hours or days.

8.4.3 Isohypses

The geopotential assumes that the Earth is perfectly flat and ideally spherical. Geopotential height is the distance above the surface of the Earth, if it is perfectly flat and in the shape of the sphere. Field of air pressure over a given area can be represented with lines where one certain isobaric surface cuts the horizontal surface with equal mutual distances (Fig. 8.5). Isohypses are showing the surface with constant pressure. For example, when looking at 850 hPa graphics, all isohypses,

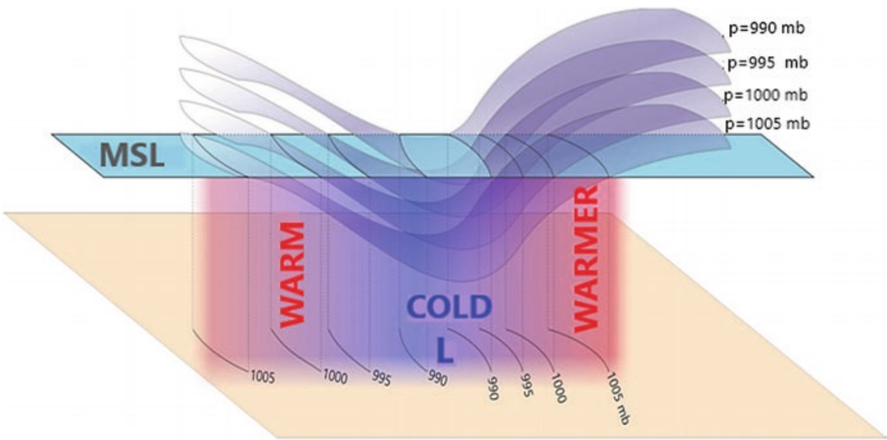


Fig. 8.4 Location of isobar surfaces in space (one section) and their intersection with mean sea level. The depictions drawn on ground-level weather maps are the projection of these cross-sections onto a horizontal plane

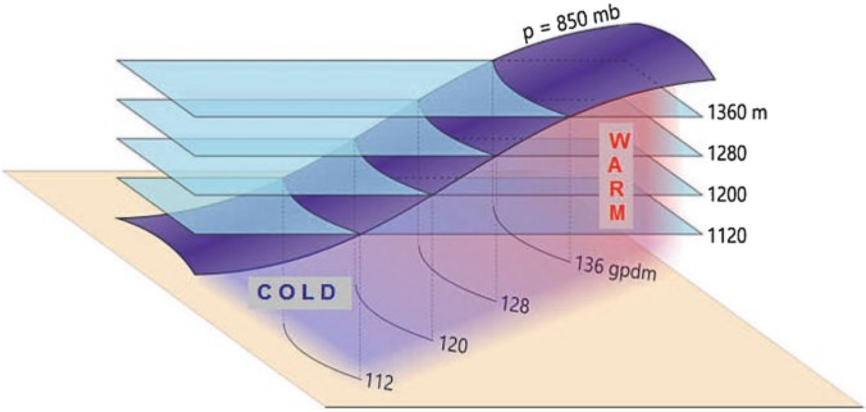


Fig. 8.5 Cross-section of isobaric surface with equidistant surfaces of equal height (geopotential surfaces). The vertical projection of these sections into the horizontal plane of the mid-sea level gives the isohypses of a given isobaric surface expressed in geopotential decametres

regardless of their value, are located at 850 hPa. Regions with low values of isohypses are correlated with low air pressure (valley), while isohypses with high values are correlated with high air pressure (ridge). Because air pressure decreases with the height of a horizontal surface, it is the greatest where the isobaric area is the largest (smallest). Such topographic maps are used to show the distribution of the field of air pressure over larger areas and height. In practice, synoptic isohypses are usually drawn every 40 geopotential metres (gpm). For the purposes of meteorology and synoptic aerology, the notion of standard isobaric area is introduced. This is an area which shows the value of uniformed pressure. Defined standard isobaric surfaces 1000, 850, 700, 500, 400, 300, 250, 200, 150, and 100 hPa, which depend on the vertical stratification of the atmosphere, are the different geopotential heights.

8.4.4 Hydrostatic Balance

Hydrostatic equilibrium implies that they are vertical in equilibrium component of the pressure gradient force and the gravity force of the Earth, and therefore there are no vertical accelerations. This condition is fulfilled for motions of meso-, synoptic, and planetary scales, so they can be considered quasi-horizontal. But where downward and upward movements occur, this balance can be disturbed. When the air layer is warm, the pressure decreases more slowly with height, which means that the distance of adjacent isobar surfaces in the vertical direction is larger (the vertical component of the pressure gradient decreases, and when the air layer is cold, it increases). Both processes in individual areas can cause the hydrostatic balance to be disturbed. Isobar surfaces in colder air will have a lower elevation and in warmer higher ones.

8.5 Pressure Systems

Pressure systems are areas that indicate a certain distribution of atmospheric pressure expressed in numerous values. High- and low-pressure systems develop as the result of different atmospheric processes (e.g. temperature differences between atmosphere, water, and land, differential heating, upper-level disturbances, solar heating or cooling). Based on a pressure systems distribution, we experience local weather in a given area. Figure 8.6 shows a graphical illustration of the general types of the pressure patterns (e.g. lows, highs, troughs, ridges). However there are another specific pressure distribution patterns such as cold and warm core lows or highs discussed below and their schematics shown in Fig. 8.7.

Low-pressure system causes weather to be generally unsettled with occurrence of clouds, winds, and precipitation that minimize daily temperature variation.

High-pressure system is associated with a stable, dry, and clear-sky weather conditions, with larger diurnal temperature changes due to increased solar radiation. The pressure readings that are taken at various weather reporting stations all over the world are transmitted to forecast offices and are plotted on specially prepared maps.

A trough is an extended region of low atmospheric pressure at the surface or aloft. A trough forms as the result of the atmospheric disturbance in the upper part of the atmosphere after cold air breakdown from higher latitudes in a warmer region creating a frontal baroclinic zone and thus increasing a jet stream that plunges the cold air towards the equator and warm air towards the poles. The near surface forces lifting of moist air behind the trough line. A trough is usually associated with clouds, thunderstorms, precipitation, winds, particularly after the passage of the trough. At the surface, lifting air under positive vorticity advection is reflected by the formation of depressions and troughs.

A ridge represents an elongated area of a high atmospheric pressure with anticyclonic curvature of wind flow. Opposite from trough, a ridge originates in the centre of a high-pressure system and extends between two low-pressure areas. Given the

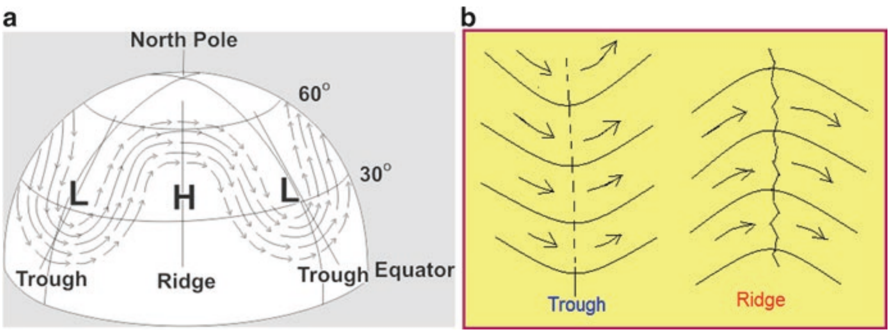


Fig. 8.6 Pressure patterns. (**a**) Lows and high at the Northern Hemisphere; (**b**) Vertical profile of a trough and ridge

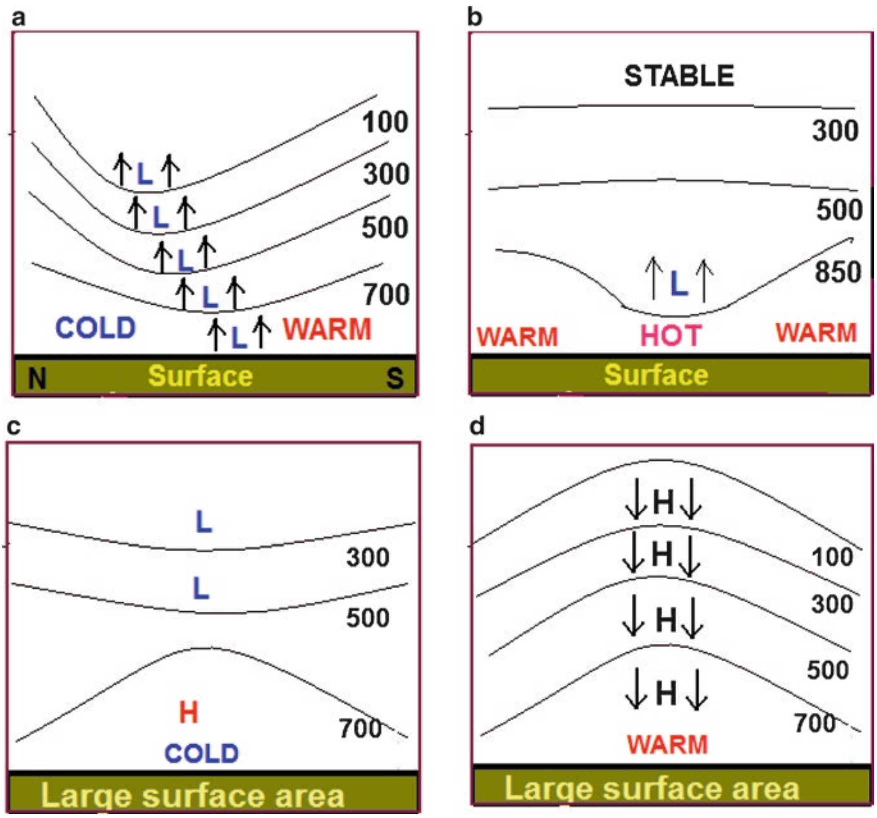


Fig. 8.7 High and low core pressure patterns: (**a**) cold core low; (**b**) warm core low; (**c**) cold core high; and (**d**) warm core high

direction of winds around anticyclonic circulation, in front of an upper ridge, the air flow from polar regions brings cold air, while behind the upper ridge line, the flow that moves from low latitudes brings mild air. The position (spot) that is associated with the maximum curvature is called the ridge line. Ridges can be recognized on a surface and upper weather charts drawn with the contour lines (isobars/isohypses), respectively, where the maximum pressure is found along the axis of the ridge. Wind convergence and the negative vorticity advection are main ingredients for development of a surface ridge ahead of the upper level ridge. The vertical downdrafts then give rise to divergence of the winds near the surface. The sinking air causes a warming in the atmospheric column and a drying, which has the effect of clearing the sky. An important atmospheric ridge is the subtropical ridge near the 30° latitude characterized by high atmospheric pressure causing blocking of western air flow associated with calm winds and stagnant air.

A shortwave trough is a mesoscale or synoptic-scale atmospheric disturbance in the mid or upper portion in the atmosphere. In general shortwave represents an embedded curve in a trough/ridge pattern that enhances baroclinity and induces upward air movement due to divergence from positive vorticity advection (PVA). It moves faster than longwave trough and influences the weather above the area of its extension and is always accompanied by precipitation.

Cold core low is pressure pattern that covers a relatively small synoptic-scale area with a rising air in all levels of atmosphere tilting to the northwest with height with low temperatures at centre of low. The air is cooled by adiabatic expansion and evaporation cooling of rain and/or snow. Most mid-latitude cyclones are cold core lows which cause widespread precipitation. Initially, cold core lows develop due to jet streaks, synoptic thermal gradient zones or vorticity spin-up that increases baroclinity and frontogenesis.

Warm core low appears as “thermal low” or “tropical cyclone”. Thermal low is shallow and most intense at surface. It usually develops over land due to strong surface heating due to positive buoyancy at the surface. In this pressure pattern, the mid and upper levels are stable with a lack of rain as the result of PBL dryness. Another type of warm core law that develops in tropical cyclones over water is deeper than thermal law, although it weakens at the upper levels. Air subsidence in centre causes compressional warming and air rising around edges of tropical cyclone eyewall. While strong cold low tilts with altitude, oppositely warm core low is strongest when there is a vertical adjustment of locations of upper- and low-level trough. A strong wind shear will weaken warm core low.

Cold core highs develop over land in high latitudes. They cover a large spatial area, with surface air sinking where polar air masses dominate with cold and dense air features. Consequently they are most intense at the surface and weakening with altitude and deepening near their source regions. When polar air mass moves into mid-latitudes, the upper-level highs will show a deep trough. Precipitation is generally lacking near cold core high centre.

Warm core highs represent deep highs with warm temperature at the surface and lack of precipitation that covers a larger spatial area and extend through large depth of atmosphere. The air subsidence generates a high pressure, despite the fact that less dense air is rising due to surface warming.

The pressure patterns on the weather map are very like contour lines on a topographic map; the high-pressure areas correspond to hills and the low-pressure areas to valleys. It is important to recognize that the pressure systems are relative to the pressure around them. A high-pressure area, in the centre of which the pressure reading is, for example, 1000 hPa (mb), is classified a high because the surrounding pressure is less than 1000 hPa. A pressure area with the same pressure reading of 1000 hPa at the centre would be classified as a low if the surrounding pressure is higher than 1000 hPa. These pressure systems are constantly moving or changing in appearance. Lows may deepen or till and highs may build or weaken. Most systems move in a general west to east direction. If we put down on a map, beside the barometre readings and wind arrows, the state of the weather, we shall see that, in the high-pressure areas, it is usually fine and clear, probably cooler, while in the low-pressure areas, it is generally rainy and cloudy on the east side and probably fine on the west side. The weather, in fact, relates to the shape of the isobars.

8.6 Daily Pressure Distribution

Air pressure has a very regular daily and annual flow in the absence of atmospheric disturbances leading to non-periodic pressure changes. The daily distribution of air pressure is characterized by two waves with a period of 12 h and thus with two maxima and two minima (Fig. 8.8). The minimum air pressure during the day occurs at 4 and 16 h and the maximum at 10 and 22 h. The daily maximum and minimum pressure (10 and 16 h) is more pronounced than the night extreme values (22 and 4 h). The daily amplitude of the air pressure decreases with latitude. It is the largest in the tropics and is several millibars (Jakarta). In temperate and higher latitudes, with stable weather, the daily amplitude is much smaller (Belgrade and Stockholm in Fig. 8.6). The daily fluctuation of pressure seems to be related to the daily fluctuation of the air temperature at the soil which is highest in the tropics and then decreases towards the north. It is also thought that a possible cause of this phenomenon is temperature fluctuations in high atmospheric layers and above all in the thermosphere, which is carried downwards. At moderate latitudes, atmospheric disturbances often occur which disrupt the regularity of the daily flow and lead to non-periodic changes in air pressure of up to 10 mb.

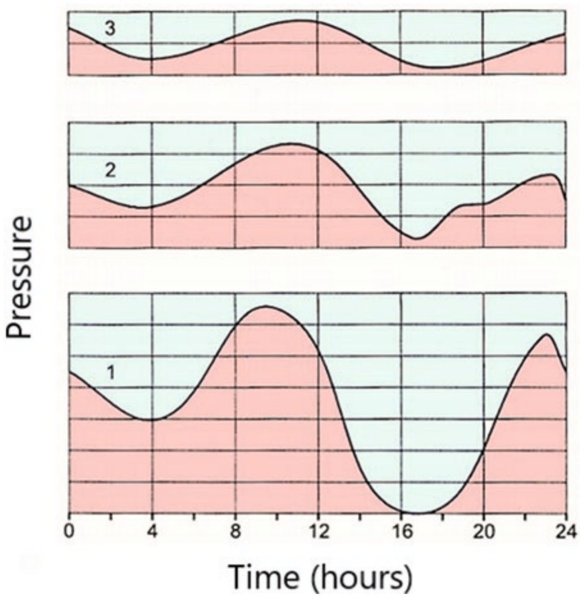


Fig. 8.8 Daily pressure distribution in *1*. Jakarta, *2*. Belgrade, and *3*. Stockholm

8.7 Reduction of the Surface Pressure to Mean Sea Level Pressure (MSL)

One important application of the static equation is to reduce the pressure at sea level. In order to be able to analyse the horizontal pressure distribution, it is necessary to reduce the atmospheric pressure values measured at the meteorological station (surface pressure) to the same reference level. The sea level was chosen as the reference level. The surface pressure measured by the mercury barometre includes instrumental repair, temperature $t = 0$ °C, and Earth's gravity acceleration at latitude $\varphi = 45°$. Special tables designed for each weather station are used to reduce the pressure at sea level. In Fig. 8.9 it is given illustration of the process of reducing the pressure to sea level.

8.8 Hydrostatic Equilibrium (Approximation)

Hydrostatics is fundamental to meteorology and other related disciplines and is applied in many different fields (Hewitt and Jackson 2003; Spiridonov 2010; Ćurić and Janc 2016). Hydrostatics offers physical explanations of the atmospheric

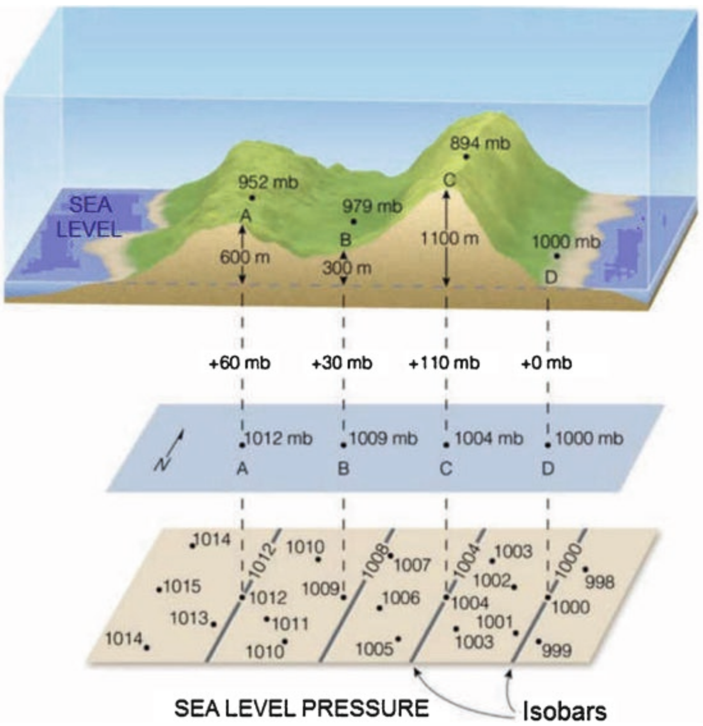


Fig. 8.9 Reduction of pressure to mean sea level (MSL)

stability and how the state variables as pressure, temperature, and density of air change with altitude. While the real atmosphere is never found in rest, under some atmospheric conditions when there is a balance between forces, almost complete calm happens. It often happens at night, when there is stable, anticyclonic weather, during the occurrence of inversion in valleys, but in the free atmosphere, where the horizontal pressure gradient is equal to zero, or in other terms when isobaric surfaces are placed horizontally. At calm atmospheric conditions, air temperature and density do not change in the horizontal direction. In atmosphere of motion, in which there are winds, it doesn't happen, and hence the difference between the atmosphere in rest and atmosphere in motion occurs. As it was mentioned earlier, the individual particle is found in the hydrostatic balance, when the vertical force of the pressure gradient is in balance with the force of gravity (Etling 2008). Since it is known that the temperature and the density of the air are related via the equation of state, we can determine how quickly the pressure will decline at a given temperature. Of course, you must assume the existence of hydrostatic atmosphere, in order to determine this relation. This is very important when it comes to atmospheric stability. There are several important elements that we need to know of, concerning the occurrence of hydrostatic balance. First, it is a good approximation of the real atmosphere in huge proportions, at the frontal zones. However, in each region where there is a significant vertical acceleration (frontal areas), hydrostatic equation does not apply when the forces set out above are not balanced. This also applies for smaller scale, such as local storms. Because the expressed turbulence cannot be assumed without hydrostatic condition, it is significantly outside of reality. In thunderstorms there is a typical movement of air lifting, and hence hydrostatic approximation is inaccurate. Hydrostatics is the science of fluids with no motion. Air pressure occurs at any height in the atmosphere, due to the force which acts on the unit area and due to the weight of air that lies above that height. Consequently, atmospheric pressure decreases with increasing the altitude above Earth's surface. The total force is directed upwards, which operates on a thin horizontal column air as a result of the decline in pressure with height, generally located in close balance with the force directed downwards, which occurs as a result of gravitational attraction. Let's derive the hydrostatic equation mathematically. The static atmosphere of the elementary particles of air is influenced by two forces: the gravitation and the pressure gradient. From the equation of motion, in this case we obtain:

$$-\frac{1}{\rho}\nabla p - G = 0 \tag{8.8}$$

i.e., in orthogonal (x, y, and z) system, we have the following three scalar equations:

$$\frac{\partial p}{\partial x} = 0, \frac{\partial p}{\partial y} = 0, \frac{\partial p}{\partial z} = -\rho g \tag{8.9}$$

The last equation is called the *hydrostatic equation*. Negative sign in the equation means that the pressure decreases with height. Equation 8.9 can be represented in another form:

$$dp = -\rho g dz \tag{8.10}$$

or as $d\Phi = gdz$ follows that:

$$dp = -\rho d\Phi \quad \text{where} \quad \Phi \quad \text{is geopotential.} \tag{8.11}$$

It is evident that in a static atmosphere, at the surface Φ = const. also surface pressure p = const. if the density ρ = const. Hydrostatic equation gives a strong relationship of air pressure by height in real (not only in the static atmosphere). Only in the areas with intense small-scale atmospheric processes (such as the cumulonimbus clouds with tornadoes, squall lines), it is necessary to take a pressure difference from that, which corresponds to the hydrostatic equilibrium. By integrating this equation from the height (z) and pressure $p(z)$, the infinite height above a fixed point on Earth, we have:

$$-\int_{p(z)}^{p(\infty)} dp = \int_{z}^{\infty} g\rho dz \tag{8.12}$$

or, since $p_\infty = 0$, it follows:

$$p(z) = \int_{z}^{\infty} g\rho dz \tag{8.13}$$

This means that the amount of pressure at height (z) is equal to the weight of air in the vertical column, a single cross-section, the area which lies above that level. If the mass of the Earth's atmosphere would be deployed in uniform globe, the pressure of sea level would be 1013 hPa, or 1.013×10^5 Pa, or an equivalent pressure of one atmosphere (1 atm). From the ideal gas law $\rho_d = \dfrac{pM_d}{RT}$ after substitution in (Eq. 8.13):

$$\frac{dp}{p} = -\frac{M_d g}{RT} \tag{8.14}$$

Assuming that T = const. and by integration of (Eq. 8.14), it is obtained:

$$\ln p\Big|_{p(0)}^{p(z)} = e^{-\frac{M_d g}{RT}} \quad \text{or} \quad p(z) = p(0)e^{-\frac{z}{H}} \tag{8.15}$$

with a scale height $H = \dfrac{RT}{M_d g} \approx 7.4\,\text{km}(T = 250\,\text{K})$.

The formula (8.17) is known as *barometric height formula* used for determining the atmospheric pressure at a certain height in the hydrostatic atmosphere (Houghton 2002; Ćurić 2000; Helmis and Nastos 2012). The pressure of any amount is equal to the weight of air above that point. The gradient of pressure is much higher near the Earth's surface, and it decreases with height. "Normal" decrease in pressure with height is observed in a standard atmosphere, which shows an idealized vertical distribution of atmospheric pressure. It is useful if the hydrostatic equation is written in terms of geopotential instead of the geometric height. Since $d\Phi = gdz$ and $\rho = p/RT$, the hydrostatic equation $gdz = -\frac{1}{\rho}\rho dp$ could be written in the form:

$$d\Phi = -\left(\frac{RT}{p}\right)dp = -RTd\ln p \tag{8.16}$$

It is evident that the change in geopotential by pressure depends only on temperature. If the former relation is integrating by height, we obtain the hypsometric formula:

$$\Phi(z_2) - \Phi(z_1) = -R\int_{p_1}^{p_2} Td\ln p \tag{8.17}$$

Meteorologists often express Φ(z) with aid of geopotential height, which is defined by $Z = \Phi(z)/g_0$, where $g_0 = 9.80665$ m/s^2 is the free average Earth's gravitational acceleration, at the mean sea level. It is seen that by numerical value, Z is almost equal to the geometric height z in the troposphere and lower stratosphere. Figure 8.10 shows the relative position of surfaces z = const and Φ = const. as a function of latitude.

Hypsometric equation can be also expressed using Z, so that:

$$Z_T = Z_2 - Z_1 = \frac{R}{g_0}\int_{p_2}^{p_1} Td\ln p \tag{8.18}$$

where Z_T represents atmospheric layer depth which is found between the surface with pressure p_1 and p_2. If the middle temperature of the layer is defined:

$$\overline{T} = \int_{p_2}^{p_1} Td\ln p\left[\int_{p_2}^{p_1} d\ln p\right]^{-1} \tag{8.19}$$

and the mean scale layer height, $H = R\overline{T}/g_0$, then the layer depth can be written:

$$Z_T \equiv \Delta Z \equiv H\ln\left(\frac{p_1}{p_2}\right). \tag{8.20}$$

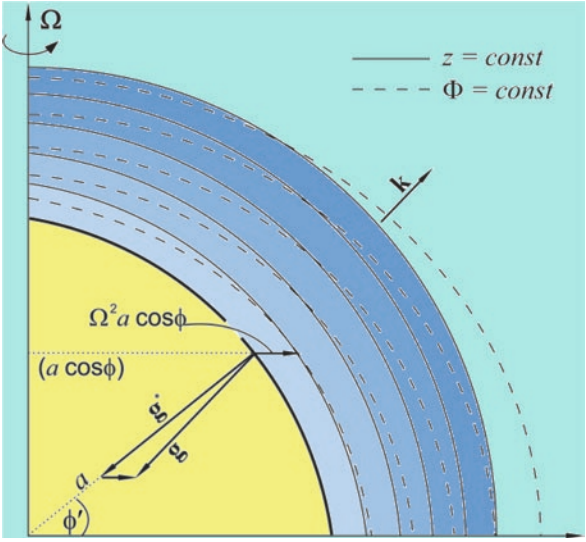


Fig. 8.10 Mutual position of z = const and Φ = const. surface as a function of latitude

It is evident that layer depth between two isobaric surfaces is proportional with the layer temperature. Hence, it is also seen that pressure is decreasing rapidly by height in cold than in warm air. We also note that in an isothermal atmosphere, the geopotential height is proportional to logarithm of relationship between pressures at height z, with the surface pressure:

$$Z \equiv -H \ln\left(\frac{p_1}{p_2}\right) \tag{8.21}$$

where p_0 is pressure at $Z = 0$. Hence it is seen that pressure decreases exponentially with geopotential height by factor $1/e$ by scale height:

$$p(z) = p(0)e^{-z/H}. \tag{8.22}$$

8.9 Standard Atmosphere

The International Standard Atmosphere (ISA) is a referent model of atmosphere which is approved by the International Organization for Standardization (ISO) as the international standard. In general ISA contains standard distribution of the pressure, temperature, density, and viscosity of the Earth's atmosphere. Other standard organizations such as the International Civil Aviation Organization (ICAO), for their specific needs, issued an additional subset to the referent model. The present standard atmosphere is adopted in 1976 by the International Civil Aviation Organization (ICAO), which succeeded US Standard Atmosphere, prepared in 1925. It assumes sea level values as follows:

- Air temperature 288.15 K (15 °C)
- Ar pressure 101,325 Pa (1013.25 mb, 760 mm of Hg, or 29.92 in. of Hg)
- Air density 1225 g m^{-2} (1.225 g L^{-1})
- Mean molar mass — 28.964 g $mole^{-1}$

For the purposes of aviation, it is important to know the exact basis of data which determine the scale of altimetre as a tool for determining the height using the atmospheric pressure. To achieve uniformity, the International Organization for Aviation Organization (ICAO) has assigned the following international standard atmosphere:

- The ICAO Standard Atmosphere, does not contain water vapor, i.e. the air is dry and chemical composition of all heights is equal.
- The acceleration of gravity is the same everywhere and it equals to (9.8062 m/s^2).
- The standard mean sea-level temperature is 15 °C, and the mean sea-level pressure is (1013.25 mb).
- At any height z (m) measured above the mean sea-level and between 0 and 11.000 m, air temperature is T = 15–0.0065 z (°C).
- For altitude above 11.000 metres, the air temperature is constant, and it equals to (−56.5 °C).

8.10 Barotropic vs Baroclinic Atmosphere

A state of the atmosphere in which air density ρ depends only upon pressure p, i.e. $\rho = \rho(p)$ is reffered to us as *barotropic atmosphere* (Hantel 2013). That is actually the state a state such that surfaces of constant pressure and constant density coincide, so that the geostrophic wind is independent of height as it is shown in Fig. 8.11a. The blue portion of the surface denotes a cold region, while the orange portion denotes a warm region. In barotropic atmosphere temperature difference is restricted to the boundary. The dotted lines enclose isobaric surfaces which remain at constant slope with increasing height. Atmospheric state in which density depends upon both temperature and pressure and in which the geostrophic wind varies with height and is related to the horizontal temperature gradient via the thermal wind equation is called baroclinic atmosphere. In baroclinic atmosphere temperature

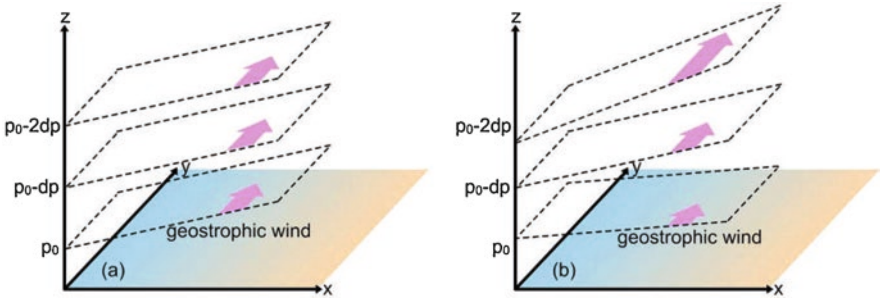


Fig. 8.11 Barotropic atmosphere (**a**) vs baroclinic atmosphere (**b**)

difference extends through the region. The isobaric surfaces increase in slope with height as it is shown in Fig. 8.11b. This causes thermal wind to occur only in a baroclinic atmosphere.

8.11 Reduction on Atmospheric Pressure to Mean Sea Level Pressure

Barometric pressure is the pressure exerted by the atmosphere at a point as a result of gravity acting upon the "column" of air that lies directly above the point. To have a consistent pressure values, meteorologists recalculate the barometric pressure measured at weather stations to mean sea level conditions. Higher elevations above sea level experience lower pressure since there is less column of atmosphere on which gravity can act. At higher altitude the air pressure decreases. Since air pressure decrease due to height above sea level is equivalent around the world, barometric pressure readings from weather stations located at any altitude can be equivalently converted into mean sea level pressure. The basic equation of statics expresses the law of change of atmospheric pressure with height in a static atmosphere. The purpose of the weather forecast is to regularly draw surface isobaric maps that show the pressure distribution above the mean sea level. The barometric height formula allows calculation of barometric pressure (p) measured at the station and reduction to the mean sea-level pressure p_0, at the mean sea level height ($z = 0$).

8.12 Stream Field

Air blows over area only when there is a pressure gradient. As it was explained above, the air will flow from a region of high pressure to a region of low pressure. The greater the difference in pressure, the faster the air flow (wind). Atmospheric air varies according to the laws of dynamics in different ways. For example, in large

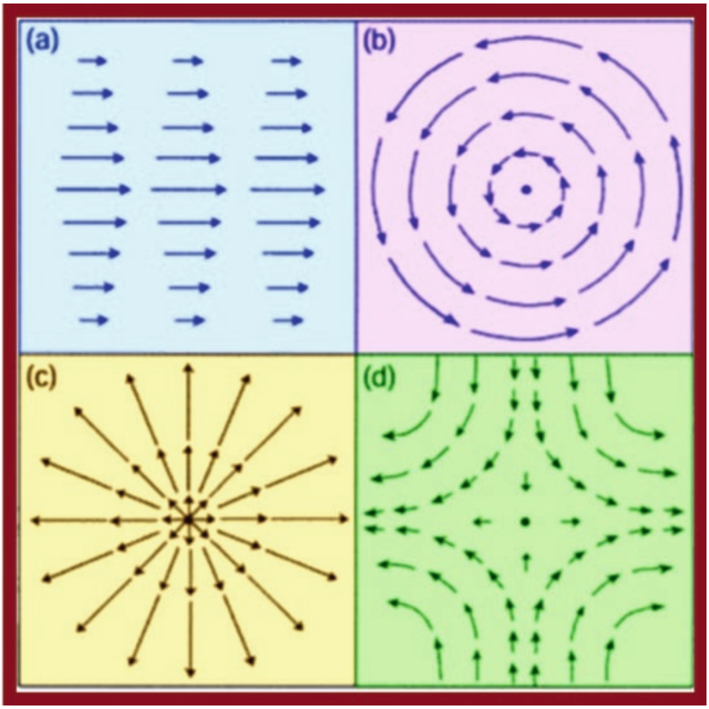


Fig. 8.12 Basic streamflows: (**a**) translation, (**b**) rotation, (**c**) divergence, and (**d**) deformation rotation

systems with an organized circulation such as cyclones and anticyclones, air rotates, i.e. it has a circular flow. At other places the air flows at a straight line; sometimes it rises and in certain atmospheric conditions, it descends; the volume of fluid increases or decreases.

Stream field. Stream field in any significant small area can be considered as a set of four basic elementary stream areas: translation, rotation, divergence, and deformation (Fig. 8.12). If a thin layer of atmosphere that surrounds Earth's surface is considered, horizontal processes are important for the movement of air masses and weather phenomena which are manifested in this layer, in relation to vertical processes. Because of this, two-dimensional elementary stream fields are briefly considered; horizontal processes are important for the movement of air masses and weather phenomena which are manifested in this layer, in relation to vertical processes. Because of this, two-dimensional elementary stream fields are briefly described.

Streamlines. Line represents the linear curve which is tangential with the local vector current speed.

Translator field. Because the speed of flow is the vector size, translator field represents movement through the displacement vector field for translation. The air moves around the same speed and direction and intensity. Streamlines that lie in the direction of flow are the parallel lines (Fig. 8.12a). That stream field does not change the volume or shape of the fluid.

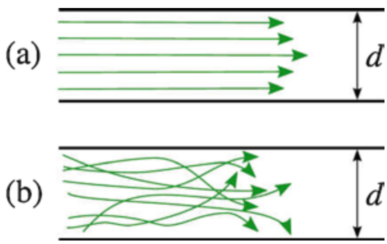
Rotation field. In this linear field, streamlines are concentric circles. The air rotates around a common centre of stream field (Fig. 8.12b). The speed of flow increases linearly with increasing distance of streamlines from the centre of rotation. During the rotation, individual air particles do not change its shape or volume.

Divergence and convergence field. Streamlines are straight and pass through the centre of the field as a joint stream point. When air flows towards the centre, flow is convergent, and that the point is called convergent point (Fig. 8.12c). If the flow is done from the center to the field, it is called the point of divergence. The convergence of an air particle does not alter its shape, while both the rate of flow and volume of particle are going to reduce the convergence point. In contrast, in the divergence, flow is from the centre towards the outer area. Then, this point is called the point of divergence. In a departure from this point, the flow increases its velocity and volume, but particles in this case do not change their shape.

Deformation field. In deformation field, streamlines are concentric equilateral hyperbolas and their velocity increases linearly in any direction from the center to the field (Fig. 8.12d). In this field, linear speed of the centre field linearly increases. Movement does not change the volume of an air particle, but it changes its shape. When approaching the common center of the hyperbola (center of the field), the dimensions of the particles decrease in the direction of flow, and when moving away from the center, the dimensions of the particles increase in the direction of flow. The axis to which air flows from one and the other side is called axis of stretching. Normally, this axis is set to tightening axis which also passes through the centre.

Laminar and Turbulent Flow The flow of air is laminar when the flow of the fluid (gas or liquid) is smoothed or regular, as opposed to turbulent flow, which is subjected to irregular deviations and mixing (Fig. 8.13). In laminar flow, sometimes known as the voltage flow, speed, pressure, and other properties of the flow in any point of the fluid remain constant. Laminar flow is presumed to consist of thin layers which are parallel to each other. When air flows at higher speeds, the flow is basically disorganized, even chaotic, and tends to form eddies. This is called turbulence or turbulent flow. In order to maintain the turbulent flow, a relatively large effect on the pressure gradient is required.

Fig. 8.13 Laminar (**a**) vs turbulent flow (**b**)



8.13 Definition of Wind

Air motion which manifests as wind is given its initial impulse by horizontal pressure gradient that results from differential heating and air density. If there is no Earth rotation and there is no friction, air flows directly from areas of high pressure towards lower pressure to transfer mass. Wind velocity is a vector quantity comprising the wind direction, where 360° is a north wind, 90° is an east wind, and so on, referring to the direction from which wind is blowing and the wind speed at a standard height of 10 m above the ground. The direction of the wind is therefore marked according to which side of the horizon air blows, and it is graphically represented with a rose in the wind. Rose in the wind has 8–16 routes. Duration of calm at the time of observation without wind is recorded in the centre of the rose. Direction is determined with the help of Wild's weathervane that is set up at the height of 6–12 cm. The speed of the wind is the flow of air particles. Wind is measured with anemometre and expressed in m/s. Wind speed depends of relief, vegetation, and other objects on the ground. The wind potential is significant for production of renewable wind energy, especially in the windy areas, as an alternate source of energy in the era of climate change (Landberg 2016).

8.14 Classification of the Winds

The winds are classified into the following groups: regular, periodic, and local winds.

8.14.1 Permanent Wind

Persistent winds are participating in the general atmospheric circulation. They are called planetary winds as they are blowing continuously through the Earth's surface. There are trade winds, antitrade winds, western winds, and eastern polar winds.

8.14.2 Trade Winds

Trade winds are surface winds blowing in the layer of the troposphere to a height of 2000 metres. They come from subtropical areas with high air pressure, 30° latitude north and south, and are directed towards the equator (Fig. 8.14). Because of the Earth's rotation, they turn to the west and thus receive a north-eastern direction (the northern hemisphere) or south direction (the southern hemisphere). Above the equator, where the trade winds from north and south collided, there is a belt of equatorial calm, which lies between 2° and 8° north latitude (Galvin 2015).

Fig. 8.14 Trade winds

Fig. 8.15 Western winds

8.14.3 Anti-trade Winds

Anti-trade winds represent upper-level winds, which are blowing over trade winds from the equator towards the subtropical areas. Under the influence of the Earth's rotation, they return to the east and thus become southwest and northwest winds, respectively. The air masses in the subtropical areas are mainly down to the Earth's surface. Due to the presence of the prevailing descendent air currents, there is no wind occurrence. Thus, the two hemispheres form a zone of subtropical calms, where clear and dry weather conditions exist. Western winds are blowing between 40° and 65° latitude (Fig. 8.15). Air masses come from the area of subtropical anti-cyclone and, because of the Earth's rotation, turn to east or west direction with the

Fig. 8.16 *Eastern* polar winds

movement. These prevailing winds are more dominant in high latitudes during winter than during summer. Eastern polar winds occur because of the air flow from the polar regions of high air pressure to depressions (low-pressure centres) around the pole (Fig. 8.16).

Under the influence of the Earth's rotation, they turn to west and east in nearly the same direction. Around 60° Latitude, they collide with western winds, lifting in height and blowing towards the poles.

8.14.4 Periodic Winds

Periodic winds are monsoons and daily winds. Seasonal changes of atmospheric pressure at the same place cause air flow over a period in one direction and during the following period in another direction. These are known as periodic winds or monsoons.

These winds are divided into winter winds, which blow from land to sea, and summer monsoons which are manifested from the sea to the mainland. The most

Fig. 8.17 Periodical winds: monsoon

famous is the East Asian summer monsoon (Clift and Plumb 2008). During this characteristic wind pattern, a humid monsoon winds can be noticed that produce heavy rains since blowing from the sea (higher air pressure) to the mainland (lower air pressure) (Fig. 8.17). Such winds blow from June to September. Winter monsoons are very dry winds blowing from the mainland (high air pressure) to the sea (low air pressure) and are manifested in the period from September to March. All winds have almost the same reason of occurrence. Differences in pressure resulting from temperature differences, caused by no equal heating of the Earth's surface, are essential factors for the occurrence of air flow into the atmosphere, called "wind".

8.14.5 Local Winds

Local winds are typical for certain areas of the Earth. Local scale circulation generally operates over 10 to 100 kilometres. Many local scale winds are created by unequal heating of the Earth. Their occurrences are related to the influence of local natural conditions and always have the same direction and support the same weather conditions. In many locations, wind patterns exist that is not easily explained by the general principles outlined above. In most cases, unusual topographic or geographic features are responsible for such winds, known as local winds.

8.14.6 Land and Sea Breezes

Winds from the land and the sea are caused by differences in daily temperature that occur between land and water (Fig. 8.18). The atmosphere is warm when spread over the land, increasing the scale height (z) and reducing the amount of the decline

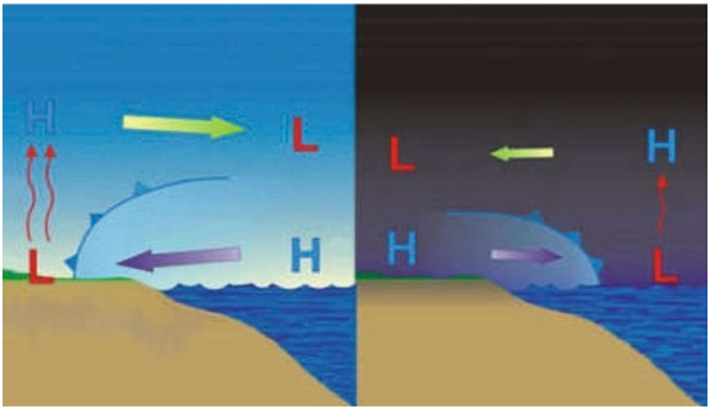


Fig. 8.18 Land and sea breezes

in pressure with height. At the height, the pressure is higher over the land than over the adjacent sea, causing a mass transport of air column above the sea. Therefore, the pressure above the sea surface increases, causing growth of the distribution of pressure, as shown in the same figure. Illustration of the sea wind shows the circulation and relative pressures in the horizontal direction, near the Earth and at the height. The Earth, unlike the sea, is heated during the day. As a result of higher temperature, the atmosphere over the land has lower density and large-scale height (z) than the atmosphere over the sea. Smaller density makes the air above the mainland suppressed, in terms of air over the sea, and it rises. Greater scale height makes the pressure higher over the land than over the sea, causing the mass to be transported from land to sea in height. The accompanying influx of mass air column above the sea increases the pressure at sea surface, placing the distribution of high and low pressure, which is shown in the picture, and it creates the shown circulation. Land or sea breezes are winds that blow from a large body of water towards or onto a landmass. They form as the result of the differences in air pressure created by differential heating capacities of water and dry land. As such, sea breezes are more localized than prevailing winds. Because water heats up and cools down more slowly than does dry land, the air along a shoreline is alternately warmer over the water and cooler over the land, and vice versa. These differences account for the fact that winds tend to blow offshore during the evening and onshore during the day.

8.14.7 Mountain and Valley Winds

The traditional components of the cycle are upslope (anabatic) winds, the daytime up valley wind, downslope (katabatic) winds, and the night-time down valley wind. In this view, each component has corresponding compensatory currents aloft, presumably to form a closed circulation.

Fig. 8.19 Mountain and valley winds

A system of diurnal winds along the axis of a valley is blowing uphill and up valley by day and downhill and down valley by night; they prevail mostly in calm, clear weather. These winds develop when the air along the mountain slopes is heated more intensely than at the same elevation above the bottom of the valley. Chinook are hot, dry winds sometimes blowing on the eastern slope of the Rocky Mountains (Fig. 8.19). In the Alps, the winds are like Chinook and are called Fen. Fen is also a strong and slope warm mountain wind. Fen brings clear and hot weather. In winter it causes avalanches and sudden melting of snow. In summer, if it is present for a long time in dry air, it may cause forest fires.

8.14.8 Katabatic Winds

This type of winds is formed when cold air, which lies above plateau area (e.g. ice sheets in Greenland and Antarctica), is set in motion under the influence of gravity. Surrounding winds associated with large urban areas, where the circulation pattern is characterized by weak wind, are blowing from the surrounding space to the city.

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Chapter 9
Atmospheric Stability

9.1 Air Stability

Atmospheric stability is a measure of atmospheric status which determines whether or not air will rise, sink, or be neutral. In general stability refers to air tendency to rise or to resist vertical motion (Salby 1996; Houghton 2002; Hewitt and Jackson 2003; Lutgens and Tarbuck 2009; Hantel 2013). As the air parcel rises, it will expand and cool adiabatically to its dew point (i.e. the level of condensation) at which clouds are formed. There are four mechanisms that trigger air rising (Markowski and Richardson 2010):

1. Orographic lifting, which occurs when air cannot move through a mountain, and it is forcing to flow over a mountain barrier.
2. Frontal lifting happens when a warmer, less dense air is forced to rise over cooler, denser air along the front line.
3. Convergence is an atmospheric condition that exists when there is a horizontal net inflow of air into the region, resulting in an upward flow. When air converges along the Earth's surface, it is forced to be raised, because it can go down.
4. Updraft, local convective rising, as the result of unstable atmosphere and vertical transport of heat and moisture through the convection (Fig. 9.1).

When air encounters an obstacle, it is forced to lift upwards across the slope. On the upwind side of obstacle, air cools adiabatically, while at the downwind side, it heats adiabatically. Hence, a more increasing humidity occurs at the upwind side than that of the downwind side. Another possible way of lifting the air is the convergence of the flow of air at Earth's surface, which forms a low pressure and increased vertical velocity, and the air begins to rise. In this case, divergence of the wind occurs in the upper layer. Occurrence of vertical (ascendant) flow of air is noted in the boundary areas, which exist in the frontal surfaces warm and cold fronts, outflow boundary surfaces (thunderstorms), and dry lines. Individual air particle may be obtained by taking convection or lifting by heating the surface (diabatic). Particles

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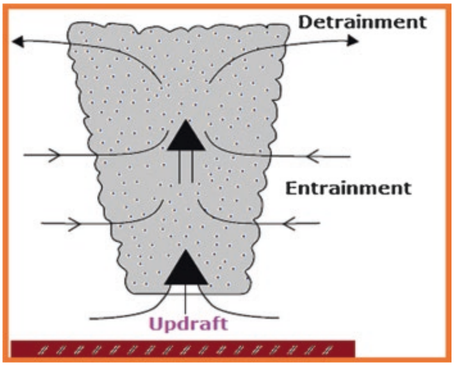


Fig. 9.1 Air rising caused by convection

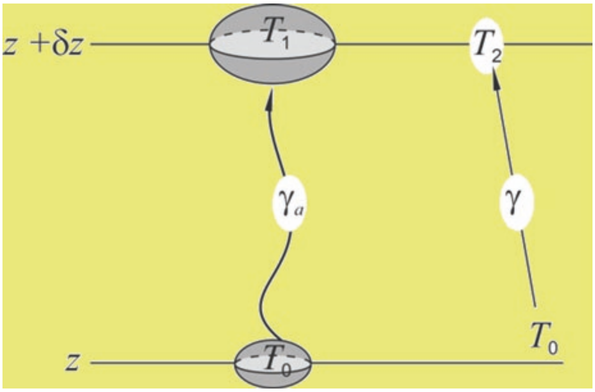


Fig. 9.2 Adiabatic and environmental vertical temperature gradient

with hot air can be collected from the surface and mixed with ambient air. The convection mechanism is responsible for development of clouds with vertical development (cumulus clouds), as precursors of thunderstorm clouds, known as cumulonimbus. The stability of the layer of the atmosphere can be estimated by comparing the vertical stratification of the atmosphere (Lutgens and Tarbuck 2009). It is obtained by upper air sounding measurements and analysis of dry and wet adiabatic rates (Fig. 9.2).

9.2 Static Atmospheric Stability

Static atmospheric stability (hereinafter referred to as atmospheric stability) is determined by comparing the vertical temperature gradient of the part of the air (or air) that is assumed to rise (or descend) dry adiabatically or moist adiabatically and the vertical temperature gradient of the surrounding air (Ćurić 2000; Ćurić and Janc 2016). Depending on the value of the temperature gradient, the atmosphere can be stable, neutral, and unstable. The value of the lapse rate is strongly dependent on the amount of water vapour in the air. Dry air cools at about 1 °C/100 m (the "dry adiabatic lapse rate"), while moist air usually cools at less than 0.6 °C/100 m ("moist adiabatic lapse rate"). The word adiabatic means that no outside heat is involved in the warming or cooling of the air parcels. To define the stability of the atmosphere, according to the vertical temperature gradient, there are three possible conditions:

Absolutely stable. The atmosphere is absolutely stable when the air at the surface is either cooler than the air aloft (an inversion) or the temperature difference between the warmer surface air and the air aloft is not very great (i.e. the environmental lapse rate is less than the moist adiabatic rate and it is positioned over the wet and dry adiabat).

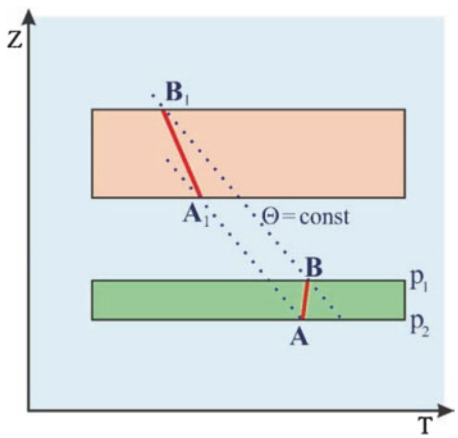
Absolutely unstable. The atmosphere is unstable when the surface air is much warmer than the air aloft (i.e., the environmental lapse rate is greater than the dry adiabatic rate, and it lies under the wet and dry adiabat).

Conditionally unstable. The atmosphere is conditionally unstable when unsaturated air can be lifted to a point where condensation occurs, and the rising air becomes warmer than the air around it. This takes place when the environmental lapse rate lies between the moist adiabatic rate and the dry adiabatic rate.

9.3 Stability Due to Air Movement

It is of great importance to know how vertical displacements of air layers affect the change in static stability. This mechanism of vertical displacement of air layers is most often due to the divergence of the ground level (divergence) in the high-pressure system (anticyclone) or due to the near surface layer air flow (convergence) in a cyclone. Observe a layer of air in the form of a square which, when lifted, remains unsaturated (Fig. 9.3). If the cross-sectional area of the air is assumed to

Fig. 9.3 The variation of air stability due to the displacement of the air layer



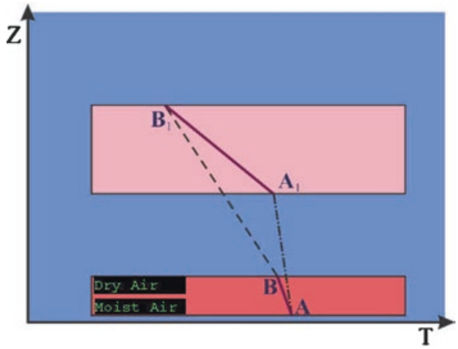


Fig. 9.4 Illustration of the mechanism of convective instability in the air layer. Line AA1 represents wet adiabat and line BB1 dry adiabat

remain unchanged, the air layer will occasionally lift to become thicker as it climbs into an area of lower pressure and less density, where the same mass of air is distributed in a larger volume. It's a layer where the air is statically stable (temperature inversion). It is evident by observing the change of temperature between points A and B on the lower and upper boundaries of the air layer, which do not change their relative position when lifting the layer. As the air layer rises, the temperatures of points A and B change dry adiabatically (lines of constant potential temperature θ) at altitude; there will be an increase in the instability in the layer, when the temperature between the points A1 and B1 declines. One can also observe the opposite case, when, e.g. in the anticyclone the air goes down. In this case, the static stability of the air increases, forming a subsidence inversion (sinking air) in that layer (Fig. 9.4). This often happens in the area of subtropical high pressure centres where significant air pollution can occur in large urban areas because inversion prevents the ventilation of air above such areas.

9.4 Convective Instability

In the mornings in summer, it often happens that the lower part of the ground layer of air (Fig. 9.4. lower position) is saturated with water vapour and in the upper part is unsaturated. Such a layer is stable, because it can also cause ground-level night inversion of temperature. Under the influence of intense heating, such a layer rises (Fig. 9.4, upper position). As it rises, the air layer becomes absolutely unstable. This occurs because the air temperature of the lower air layers changes moist adiabatically, by line AA1, which is slower than the temperature rate of the upper air layers. Air temperature of the upper-unsaturated air layers changes dry adiabatically, along line BB1. In such cases, the rising air layer is said to be convectively unstable. The mechanism described is responsible for the appearance of the most developed convective clouds, cumulonimbus, in the tropics and temperate latitudes. This mechanism of instability apparently occurs in the lower layers of the troposphere.

9.5 Low-Level Inversions

Inversions play an important role in determining cloud forms, precipitation, and visibility. An inversion acts as a cap on the upward movement of air from the layers below. As a result, convection produced by the heating of air from below is limited to levels below the inversion. A weather situation typical for occurrence of low-level temperature inversion is shown in Fig. 9.5.

During winter period at the Northern Hemisphere (or summer period at Southern Hemisphere), the solar radiation is weaker, so less heating comes to the Earth's surface. In addition, if there is a snow cover, the ground becomes colder due to the process of radiation cooling, and rapid heat and near surface temperature decrease. The persistent high-pressure system with stable atmospheric conditions, mainly calm or weak horizontal winds, suppresses air mixing near the surface, and clear skies increase the rate of cooling at the Earth's surface. The absence of vertical and horizontal mixing near the Earth's surface favours the development of a low-level temperature inversion, where the positive vertical temperature gradient exists in each atmospheric layer. In such stable atmospheric stratification, the portion of the column atmosphere acts as an energy barrier that does not allow vertical transport and mixing of ground air with the environment. The specific topography of the terrain (the landscape) especially in urban areas and valleys plays also significant role in the formation and intensity of temperature inversion. During the day, surface inversions normally weaken or disappear when the Sun warms the ground. Inversions play an important role in determining cloud forms, precipitation, and visibility. An inversion acts as a cap on the upward movement of air from the layers below. As a result, convection produced by the heating of air from below is limited to levels

Fig. 9.5 Specific weather situation during temperature inversion

below the inversion. A weather situation typical for occurrence of low-level temperature inversion is shown in Fig. 8.5. The formation of low-level inversions and their persistence for a longer period accelerates the pollution accumulation and contributes to formation of extreme pollution episodes with a poor air quality (Spiridonov 2010). There are four types of inversions: radiation, turbulence, subsidence, and frontal.

Radiation (night-time) or ground inversion develops when air is cooled by contact with a colder surface until it becomes cooler than the overlying atmosphere; this occurs most often on clear nights, when the Earth's surface cools off rapidly by radiation (see Fig. 8.4). If the temperature of surface air drops below its dew point, fog may result. Topography largely affects the magnitude of ground inversions. If the land is rolling or hilly, the cold air formed on the higher land surfaces tends to drain into the hollows, producing a larger and thicker inversion above low ground and little or none above higher elevations.

A turbulence inversion often forms in lower layer of the planetary boundary layer (PBL) at the near surface layer, when calm air overlies turbulent air. Within the turbulent layer, vertical mixing carries heat downwards and cools the upper part of the layer. The unmixed air above is not cooled and eventually is warmer than the air below; an inversion then exists.

A frontal inversion occurs when a cold air mass undercuts a warm air mass and lifts it aloft; the front between the two air masses then has warm air above and cold air below. This kind of inversion has considerable slope, whereas other inversions are nearly horizontal. In addition, humidity may be high, and clouds may be present immediately above it.

A subsidence inversion develops when a widespread layer of air is sinking. The layer is compressed and heated by the resulting increase in atmospheric pressure, and as a result the lapse rate of temperature is reduced. If the air mass sinks low enough, the air at higher altitudes becomes warmer than at lower altitudes, producing a temperature inversion. Slow descent air in areas with a high pressure is an important factor in the modification of air mass. This slow descent air is responsible for the development of inversions formed in the free atmosphere – a layer lying high above the Earth's surface. These inversions of subsidence form slow descent air that is heated by adiabatic compression. Air subsidence almost never goes down below the Earth's surface. Near the Earth's surface, there is always some weak turbulent mixing. This poorly downs the air intercepts by turbulent mixing.

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Chapter 10
Atmospheric Moisture

10.1 Water Vapour in the Atmosphere

Water vapour is present in the atmosphere primarily due to the evaporation of water from the Earth's surface. In the atmosphere, it is distributed by large-scale atmospheric disturbances, convection, turbulent mixing, and diffusion. The water vapour content decreases rapidly from the ground moving upwards into the atmosphere. It declines faster in the free atmosphere than above the mountain slopes, because in this case, water vapour is in direct contact with the evaporating soil. In contrast, in temperature inversions, the water vapour content can increase with height. The air is moist due to the presence of water vapour. Humidity is one of the most important meteorological elements. The water vapour content of the atmosphere is quantitatively represented by the quantities that determine the amount of water vapour in the air, as well as those that show the degree of saturation of water vapour in the air.

Moisture processes in the atmosphere have a significant importance in hydrological cycle, the atmospheric boundary layer processes, condensation, evaporation, formation of clouds and precipitation, and Earth's energy balance (Smith 1996; Pruppacher and Klettt 2004; Anthes 1996; Straka 2009; Allaby 2007). Water cycle driven by solar radiation is called hydrological cycle (see Fig. 10.1). It is a continuous movement of water from the Earth's surface to the atmosphere and back to the surface.

This water cycle encompasses the atmosphere, Earth's surface, surface water, and groundwater (Spiridonov 2010; Saha 2008). As water moves through this cycle, it changes its aggregate state into liquid, solid, and gas. The cycle shows a steady water flow from the ocean to the atmosphere, from the atmosphere to land, and from land back to the seas. Basic water cycle involves the physical processes (e.g. evaporation, transpiration, evapotranspiration, condensation, precipitation, infiltration, and runoff). The quantitative view of the hydrological cycle is known as water balance. The quantitative measure of the amount of humidity found in air varies because of several factors. Two important factors are evaporation and condensation.

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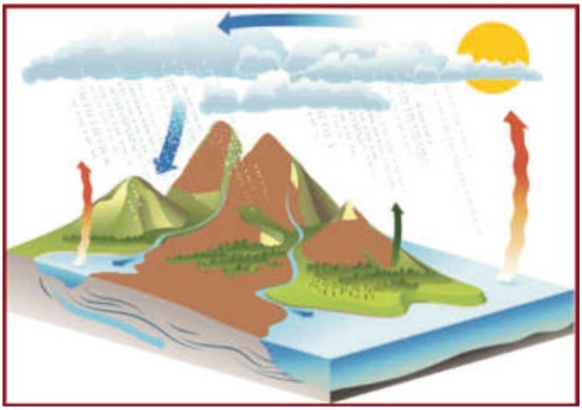

Fig. 10.1 Hydrological cycle

Humidity indicates a general term used to describe the amount of water vapour in the air.

10.2 Air Moisture Quantities

The amount of water vapour in the air is determined through absolute humidity, specific humidity, mixing ratio, and water vapour pressure.

The *absolute humidity* or water vapour density is defined as the mass of water vapour in a unit of air volume. It is most commonly denoted with (a), expressed in gm^{-3} or kgm^{-3}, and it is not often used to characterize humidity.

Specific humidity (q) is defined as the mass of water vapour in unit mass of moist air. Although it is a dimensionless quantity, it is most commonly expressed in g/kg. The mixing ratio is defined as the mass of water vapour in a unit mass of dry air. Both quantities change in the individual air parcel, if water vapour or dry air is added or removed from it.

The *mixing ratio* is defined as the mass of water vapour in a unit of mass of dry air. Both sizes change in a fraction of the air if water vapour or dry air is added to or subtracted from it.

Water vapour pressure, (e), is the partial pressure in any gas mixture that is found in thermodynamic equilibrium with liquid water or with its condensate state. As water vapour in the air increases, so does its pressure. It is expressed in pressure units, usually in millibars. The actual water vapour pressure is not measured by direct measurements but is calculated based on indirect measurements by a psychometric method, using dry and wet-bulb temperature data. The *maximum water vapour pressure*, (e_w), is the saturated water vapour pressure. The value of this pressure indicates the total amount of water vapour in the air at a given temperature. Water vapour saturation occurs in that moment when saturated vapour pressure of a system has attained equilibrium with its condensate. More specifically, the maximum water vapour pressure over the flat surface of pure water is discussed here. The maximum water vapour pressure is only a function of temperature (Figs. 10.2 and 10.3). At a temperature of $T = 0^{\circ}C$, the maximum water vapour pressure is $e_w = 6, 11\ mb$. Below $T = 0^{\circ}C$, water vapour pressure in respect to ice is less than that in relation to water. Their difference is greatest at $T = -12^{\circ}C$ and is approximately 0, 28 mb (Fig. 10.2). Many relations are used in practice to calculate the maximum water vapour pressure of varying degrees of accuracy. One is the Tetens formula:

$$e_w = 6{,}1078e^{17{,}269388x(T-273.16)/(T-35{,}86)} \tag{10.1}$$

The relation between specific moisture (mixing ratio) and water vapour pressure is:

$$q = \frac{m_v}{m_v + m_s} \tag{10.2}$$

where m_v and m_s are water vapour masses of dry and moist air. The specific volume could be expressed with aim of a water vapour density for dry air:

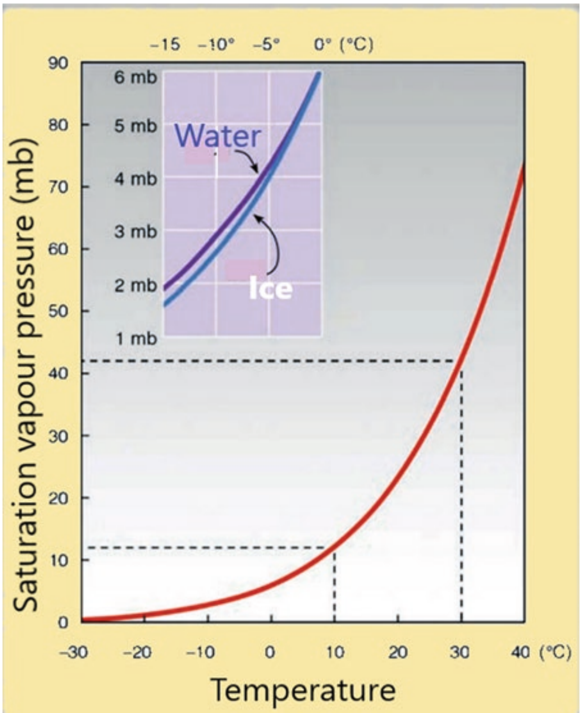

Fig. 10.2 The maximum water vapour pressure over flat surface of pure water (ice) as a function of temperature

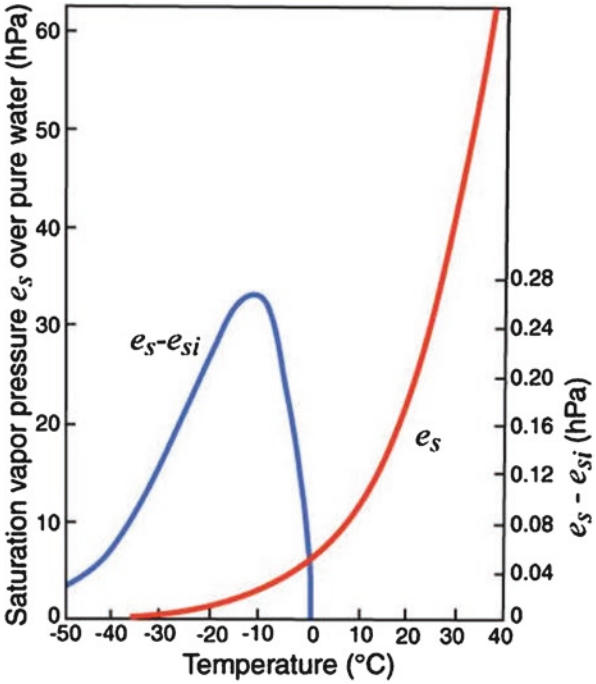

Fig. 10.3 The difference between the maximum water vapour pressure in respect to water and ice, and the same in respect to water, as a function of temperature

$$q = \frac{\rho_v}{\rho_v + \rho_s} \tag{10.3}$$

By making use of the equation of state for a dry air and water vapour, and Dalton's law, the following relation is obtained for the specific moisture:

$$q = 0{,}622 \frac{e}{p - 0{,}378e} \tag{10.4}$$

and the mixing ratio (r):

$$r = 0{,}622 \frac{e}{p_s} \tag{10.5}$$

The quantities which determine the water vapour saturation rate are the relative humidity, moisture deficit, dew point temperature, and wet-bulb temperature.

10.3 The Relative Humidity

Actually, the relative humidity is a measure of the amount of water vapour in the air, compared to the maximum amount of water vapour, expressed in (%) percentages. Relative humidity according to the American Society Glossary (AMS Glossary) represents the ratio of the vapour pressure to the saturation vapour pressure with respect to water. This quantity is alternatively defined by the World Meteorological Organization (WMO) as the ratio of the mixing ratio to the saturation mixing ratio. These two definitions yield almost identical numerical values. Relative humidity is usually expressed in percents and can be computed from psychrometric data. When the air cannot absorb the additional amount of moisture (when fully saturated), its relative humidity is around 100%. Relative humidity can be changed in two ways (Ćurić 2001):

1. By changing the amount of moisture in the air
2. By change in air temperature

With the addition of moisture in the air at a constant temperature, relative humidity increases. With the removal of moisture, relative humidity decreases. When the water vapour content remains constant, air cooling results in an increase of relative humidity. Oppositely, air warming decreases its relative humidity. There are three ways in which the change of temperature causes additional changes in relative humidity:

1. *Daily (versus night-time) changes in temperature*
2. *Temperature changes as a result of air moving horizontally from one to another place*
3. *Changes caused by vertical movement of air in the atmosphere*

Relative humidity (U) is defined as the ratio of actual and maximum water vapour pressure expressed as a percentage:

$$U = \frac{e}{e_w} x100 \tag{10.6}$$

Actual and the maximal vapour pressure could also be expressed through the equation of state for water vapour as:

$$e = \rho_v R_v T_v, e_w = \rho_w R_w T_w \tag{10.7}$$

where ρ_v and ρ_w are corresponding water vapour density and maximum water vapour density in the air at given temperature. Thus, the relative humidity is given in the following form:

$$U = \frac{\rho_v}{\rho_w} x100; U = \frac{r}{r_w} x100, \tag{10.8}$$

Relative humidity is most commonly used to describe humidity. One responds to changes in the relative humidity of the air rather than changes in the content of water vapour in the air. When the relative humidity is low at high temperatures, the person sweats and the water evaporate from the skin, subtracting heat in the amount of latent evaporation heat mainly from the human body and partly from the surrounding air. One then has a subjective feeling of being temperature air lower than it is. At high air temperatures and high relative humidity values, the person sweats, and the sweating persist on the skin because its evaporation is less and thus the cooling of the air with the body is less. Daily and annual changes in water vapour pressure and relative humidity of air depend on the surface and air temperature, since the intensity of evaporation depends on these meteorological parameters. Evaporation also depends on the wind speed, so the analysis of daily and annual changes can be most easily described in the case of stable weather, when the observed area is not affected by any atmospheric disturbance. Above water surfaces or lands that accumulate a lot of water available for evaporation, one maximum and one minimum of water vapour pressure occur during the day. It follows the daily distribution of air temperature. In winter, a similar flow of water vapour pressure can occur above land, because of the low temperatures and small amounts of water vapour are enough for saturation. In the warmer months of the year, above the land surface, water vapour pressure daily has two maximums, at 9 and 9 pm, as well as two minimums, at 4 and 3 pm. First, the main daily maximum comes from the warming of the soil after sunrise when the water from the surface evaporates more rapidly. The second maximum is related to the night characterized by weak vertical movements, while the soil is still warm and allows evaporation. The main minimum occurs before sunrise, because then the soil is coldest due to terrestrial radiation. As a result, water vapour condensation on the ground can sometimes occur. The second minimum occurs in the afternoon shortly after the maximum of the soil temperature, as the pronounced convective and turbulent motions transport water vapour from the ground upwards. In mountainous areas, the daily flow of water vapour pressure has only one maximum (in the warmest hours) and one minimum (at night) because during the day, the valley wind transports additional amounts of water vapour to higher altitudes. By contrast, at night the mountain breeze drives water vapour towards the valley. Relative humidity generally has an opposite daily flow of air temperature as the maximum water vapour pressure decreases with decreasing temperature. Therefore, relative humidity reaches a maximum before sunrise and a minimum in the afternoon, between 3 pm and 4 pm local time. The daily flow of relative humidity above the mountain peaks is completely opposite. That is where the minimum occurs in the early morning, a maximum in the afternoon, when convective movements are most pronounced. This daily flow is caused by the mountain and valley winds that alternate during the day. Unlike the air above land, the relative humidity changes very little during the day over large water areas due to the small daily fluctuation of the surface water temperature and high evaporation. Going from the pole to the Equator, the average relative humidity has highs in the tropics and near the poles, due to low temperatures and low water vapour content. The minimum relative humidity occurs at latitudes of 30 ° due to vast desert areas.

10.4 The Moisture Deficit

The moisture deficit D is defined as a difference between the maximum and actual water vapour pressure and is expressed in mb, i.e.

$$D = e_w - e. \tag{10.9}$$

The dew point, T_d, is the temperature at which the air must be cooled to reach saturation, without changing the water vapour content and atmospheric pressure. Then the actual water vapour pressure becomes equal to the saturated water vapour pressure. When the ground air cools during quiet and clear nights and at sufficiently high relative humidity, dew or fog forms on the ground. In such situations, further cooling of the air slows down as latent heat is released by condensation of water vapour.

10.5 A Dew Point Temperature

The dew point is equal to the air temperature only if the air is saturated with water vapour and in all other cases, it is lower. It has great application in meteorology. For example, it is used when forecasting frost. If the dew point is above = $0^{\circ}C$, the risk of frost is low. At temperatures below $T = 0^{\circ}C$, a frost point is defined in addition to the dew point temperature. It is analogously defined as the dew point by observing the pressure of saturated water vapour with respect to ice. Frost is formed by the deposition of water vapour at temperatures below = $0^{\circ}C$. The dew point is lower than the frost point, because at the same temperature, the maximum water vapour pressure with respect to ice is lower than that with respect to water (Fig. 10.3). By definition, frost is not produced by freezing dew. Sometimes it happens that dew first forms and the droplets freeze upon further cooling. This is called frozen or *white dew*. The wet-bulb temperature, T_w, is the temperature that a particle of air would have if it had evaporated water until the water vapour became saturated, whereby the latent heat required to evaporate the water is taken from the particle. Using the first principle of thermodynamics, it can be written as:

$$\left(m_s + m_v\right) c_p \left(T - T_w\right) = L\left[m_w\left(T_w\right) - m_v\right], \tag{10.10}$$

where m_w is the mass of saturated water vapour at a wet-bulb temperature. The wet-bulb temperature is higher than the dew point and less than the dry-bulb temperature for unsaturated water vapour. It is higher than the dew point because water vapour saturation is achieved at its higher content, which corresponds to higher temperature (Fig. 10.3).

10.6 The Phase Changes

Even in natural conditions, water can be found in all three aggregates (Fig. 10.4). The diagram is divided into three areas with the corresponding aggregate state of water: ice, water, and vapour. These areas are separated by lines along which the two-aggregate water (phase) states are in equilibrium. The Earth's atmosphere contains water vapour which is found in gaseous state. It is a variable, invisible, odourless, gaseous constituent in the atmosphere that can change its aggregate state (liquid, solid, and gas). Processes which contribute in changing the state of water vapour are referred to us as *phase changes*:

- Evaporation (liquid to gas)
- Condensation (gas to liquid)
- Melting (solid to liquid substance)
- Freezing (liquid to solid matter)
- Sublimation (solid to gas matter)
- Deposition (gas to solid)

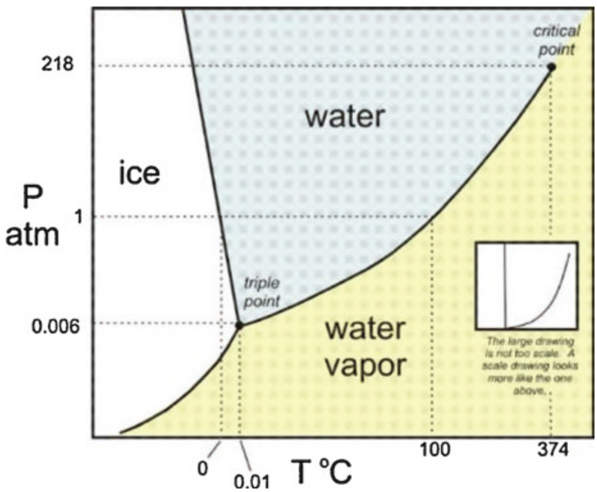

Fig. 10.4 Phase diagram of water

Water vapour is extremely important to the weather and climate. Without water vapour, there would be no formation of clouds and precipitation. All the water vapour that evaporates from the surface of the Earth eventually returns as precipitation – rain or snow. Water vapour is also the Earth's most important greenhouse gas, accounting for about 90% of the Earth's natural greenhouse effect, which helps keep the Earth warm enough to support life. When liquid water is evaporated to form water vapour, heat is absorbed. This helps to cool the surface of the Earth. This "latent heat of condensation" is released again when the water vapour condenses to form cloud water. This source of heat helps drive the updrafts in clouds and precipitation systems, which then causes even more water vapour to condense into cloud and more cloud water and ice to form precipitation. The phase diagram of water is presented in Fig. 10.4. It shows the dependence of water vapour pressure from the temperature. When it is separated, the three phases can be noticed, solid state of water, liquid, and gas. The diagram shows that at low temperature (solid state), the ice is in a stable condition. With moderate temperatures and high water vapour pressures, the liquid state of water in the stable phase at high temperature and low water vapour pressure (gas condition) is vapour found in a stable phase. The sublimation curve distinguishes a solid from gas state of water. This line shows the dependence of water vapour pressure from ice as a function of temperature. The axes correspond to the pressure and temperature. The phase diagram shows, in pressure-temperature space, the lines of equilibrium or phase boundaries between the three phases of solid, liquid, and gas. At the interface between air and liquid water, water molecules are either evaporating, which means changing phase from liquid to gas, or condensing as changing phase from gas to liquid. When the vapour pressure in the atmosphere is in equilibrium with the vapour pressure of a water or ice surface, there is no net exchange of water molecules in both directions, and the atmosphere is said to be saturated. When liquid water condenses, more water molecules are changing phase from gas to liquid than changing phase from liquid to gas. If the rate of evaporation is lower relative to the rate of condensation, the air is said to be supersaturated with respect to water vapour. The amount of liquid water doesn't change. This condition doesn't last too long. Oppositely, when liquid water evaporates, in that case more water molecules are changing phase from liquid to gas than changing phase from gas to liquid. If the rate of evaporation is lower than rate of condensation, the air is said to be unsaturated with respect to water vapour. The amount of liquid water decreases.

10.7 Condensation and Evaporation

The process of condensation occurs when water vapour is transformed into liquid. For condensation to occur, air must be saturated with water vapour, and there must be a presence of surface where it could condensate. Additionally, at the ground layer of the atmosphere, small (minor) hygroscopic (water absorbent) particles are present, known as condensation cores (nucleus), which serve as surfaces on which water vapour can condense. The evaporation is a reverse process of condensation of a water vapour. Water evaporation is a significant process in the atmosphere and is one of the basic components of the water cycle in nature (commonly referred to as the hydrological cycle) as well as the heat balance (via the latent evaporation heat flux). In meteorology, evaporation is expressed in mm of evaporated water over a period (1 mm of evaporated water corresponds to 1 l of water evaporating from an area of 1 m^2). The intensity or rate of evaporation of water is defined as the amount of water that evaporates from a unit surface in a unit of time. The evaporation of water on the Earth's surface takes place from the free surface water in liquid and solid (ice and snow) or from the ground, including vegetation. This complex process depends on the temperature of the evaporating surface, the saturation of water vapour in the air above the evaporating surface, the horizontal and vertical movements of the air above the evaporating surface, the air pressure, the amount of the precipitation, the relief and the position of the place towards the sides of the world, the proximity of groundwater to the soil surface, as well as the type and condition of the plant cover. In the discussion that follows, only the impact of the first four factors will be considered. The evaporation intensity increases as the surface temperature, where evaporation occurs, increases. The velocity of water molecules increases with increasing temperature, so fewer or more of them can leave the water surface. Evaporation is higher if the saturation of water vapour in the air is lower than the surface from which the evaporation takes place. As stated earlier, air can only contain a certain number of waters vapour molecules at a given temperature. When the air temperature is higher, a larger number of water vapour molecules are required to achieve saturation. This means that evaporation also depends on the temperature of the air above the evaporating surface. Evaporation only takes place until saturation of water vapour in the air is reached. In the atmosphere, saturated water vapour can quickly become unsaturated due to air mixing by diffusion as well as horizontal and vertical air movements. Evaporation increases with increasing wind velocity as it drives water vapour above the evaporating surface. This is primarily due to the dry wind that brings in less humid air. When the wind brings in moist air, its effect on evaporation is much less. Convection can also influence evaporation, as the humid air from the ground floor is transported upwards by updrafts, while the dry air is released from the higher layers into the lower layers by downdrafts. In addition to the strength of the wind, changing its speed is also important for the evaporation process with height. The greater the change in wind speed with altitude, the more intense the turbulent mixing of the air in the vertical direction and thus the evaporation. Evaporation decreases with increasing air pressure due to the slowing of the molecular diffusion process. This process is the least significant of all the processes that transport water vapour in the atmosphere. Water evaporates faster, under the same conditions, from the surface of fresh water than saltwater, due to stronger adhesion forces between water molecules and salts. Daily and annual evaporation distributions are aligned with the corresponding ones by the temperature of the evaporating surface and the air above. During the day, evaporation is greatest between 12 am and 4 pm local time. Then the lowest saturation of water vapour in the air is due to the highest temperatures of the surface from which evaporation

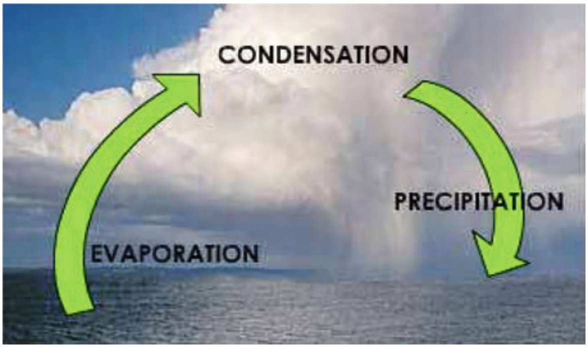

Fig. 10.5 The process of condensation and evaporation

takes place as well as ground air, while the convective movements are the most intense. The evaporation minimum is before sunset when due to terrestrial radiation of soil, condensation or deposition of water vapour and the formation of dew or frost on the soil surface often occur. The annual evaporation rate is also consistent with the annual temperature distribution. The evaporation of water from the ocean or a moist surface requires an energy source and wind to remove the evaporated water vapour. The rate of evaporation also depends on the vertical gradient of vapour pressure in the planetary boundary layer (PBL). The latent heat of vaporization required to convert liquid to vapour is about 2.5 x 10^6kg^{-1}. As an illustration, the energy needed for evaporation of a given mass of liquid water is about 600 times relative to the energy required to increase its temperature by 1 °C and up to 2400 times the energy required to increase the air temperature by 1 °C. Both processes, condensation and evaporation, are illustrated in Fig. 10.5. To begin the process of evaporation, favourable atmospheric conditions, higher air temperature, and low water vapour saturation pressure should exist. Wind and favourable topography of the terrain accelerate the process of evaporation of water vapour.

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Chapter 11
Clouds and Precipitation

Clouds are a visible manifestation of condensation or deposition in the atmosphere comprising suspended water droplets and ice crystals (Wang 2013). They exist in a wide variety of forms from sheets to rolls, to vertical towers in convective clouds, and they occur from the surface in the form of fog to the upper troposphere ~10–15 km. Clouds have the ability to reflect and also with a small amount to absorb the solar radiation. The cloud reflection (albedo) is high and depends on the cloud optical depth and the liquid water content. The optical depth represents a measure of absorbed or scattered solar radiation along a path through the cloud, i.e. the cloud transparency. On average, clouds cover at about 66% of the Earth's surface and, with high albedo, represent a significant factor in reducing the incoming solar radiation. In addition, cloud water droplets also have such absorption ability but with lower percentage amount.

11.1 Formation of Clouds

Prior to formation of clouds and precipitation, certain physical processes in the atmosphere occur. The first such process is a process in which a part of the atmosphere with a relatively large volume is increasing relative humidity up to saturation, i.e. around 100%. It usually happens because of the vertical movements of the atmosphere when it is unstable and stratified, or because of the dynamic and thermal processes that occur in it. The other process is the transformation of various categories of water in small scale: nucleation, diffusion, and accumulation, and it is the subject of study in a specific scientific topic called microphysics of clouds (Ćurić 2001b; Pruppacher and Klett 2010; Straka 2009).

Cloud condensation level (CCL) is the altitude (level) at which air cools dry adiabatically until it reaches saturation. At this level, the water vapour condenses by activation on small particles suspended in the air called cloud condensation nuclei (CCN) and clouds form. When the individual air parcel cools to the dew point, i.e.

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when the ratio of mixture of the air parcel and the mixture ratio of saturation are equal, the relative humidity increases around 100%. If rising continues, the particle rises wet adiabatically (forming a cloud). As it was described in Sect. 10.4, the absolute humidity is the measure of water vapour (moisture) content in the air, regardless of temperature. On the contrary, the occurrence of clouds mainly depends on the relative humidity, which quantitatively measures the degree of saturation and thus the likelihood of condensation (see Wang 2013). Another important detail is that about 99% of the water vapour is present in the troposphere, while less than 1% resides in the stratosphere. Hence, it is not surprising that almost all types of clouds from small cumulus to large cumulonimbus (Cb) clouds occur entirely within the troposphere (Fig. 11.1). Even the clouds of very strong storm systems, such as hurricanes and tornadoes, are basically confined within the troposphere.

Clouds form wherever air is cooled below its dew point, whether by radiation, by mixing with cooler air, or by ascent in the atmosphere with resultant decompression. The amount of water vapour, which can exist in each volume in equilibrium with a plane surface of pure water, is a function of temperature only (Heintzenberg and Charlson 2009). Air containing this amount of water vapour is said to be saturated. Any water vapour in excess of the amount required for saturation is theoretically available for the formation of a water cloud.

The rate at which water vapour is made available for the formation of cloud droplets by a known rate of cooling can be calculated from the Clausius-Clapeyron equation, which relates the saturation vapour pressure to the temperature in degrees Kelvin. The Clausius-Clapeyron equation, allows the content of water vapor to be determined for the air parcel that is cooling, as well as the amount of increase in water or ice saturation. When air rises, it cools where condensation or air saturation with water vapour occurs. Condensation depends on the moisture content in the air that rises vertically. Moist air requires less cooling, and it rises to the level of saturation. Dry air requires more cooling and more lifting to reach the point of saturation. Two important processes take place in the atmosphere before the formation of

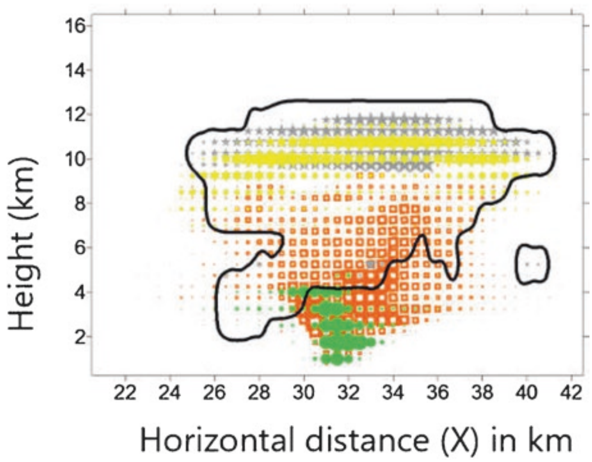


Fig. 11.1 Numerical simulation of Cb cloud with formation of hydrometeors

clouds and precipitation occurs. The first process is when in a larger volume of atmosphere, relative humidity increases to saturation, i.e. 100%. In principle, this is due to vertical air movements or as a result of some dynamic and thermodynamic processes. The second process includes three basic processes: nucleation, diffusion, and collection. These processes take place on a very small scale which are proportional to the dimensions of the cloud elements (cloud drops and ice crystals) and precipitation elements (raindrops, snow, and hail). The processes of the second type are those responsible for formation of cloud elements and take place in their further growth. They are mutually connected and highly dependent on the dynamic characteristics of the environment.

11.1.1 Air Saturation Mechanism

Before detail description of the necessary conditions for formation of cloud elements, we will briefly recall the basic factors which contribute to the change of the relative humidity of air (Khvorostyanov and Curry 2014). This can be seen from the differential form of relative humidity expression U:

$$\frac{dU}{U} = \frac{dq}{q} - A\frac{dT}{T} + \frac{dP}{P} \tag{11.1}$$

where q is the specific moisture, T the temperature, P the air pressure, and A the ratio of total and external heat of evaporation of water. Air cooling occurs as the result of heat removal (diabatic cooling), mixing of warmer and cooler air, and adiabatic cooling. The following processes lead to increase of the relative humidity:

1. Water vapour supply from different water sources: oceans, seas, lakes, and rivers.
2. Lowering the temperature. Thus, cold air is closer to saturation at the same water vapour content.
3. Increasing the air pressure. The increase in pressure at constant temperature leads to decrease of air volume, or water vapour compression on a smaller space, which means density increases or consequently the relative humidity increases.

The term for differential change of pressure is one order magnitude smaller than the other terms. The rate of the third term slightly increases with height (due to the pressure decrease). Clouds usually form in areas with dropping pressure (pressure decrease), which means the term which denotes the change of pressure makes the environment unfavourable for formation of clouds.

11.1.2 Adiabatic Cooling

When the air expands adiabatically, it cools, approaching to saturation. As the pressure decreases with height (see Eq. 11.1.), it leads to decrease in air saturation. In other words, in order for the condensation process to take place under such conditions, the cooling in vertical motion should be larger than when the expansion occurs at the constant air pressure. Another mechanism for condensation to occur in the updraft motion is that the vapour pressure of saturated air needs to decrease more rapidly than the total pressure (or water vapour pressure). That is seen from the differential form of the "first law of thermodynamics" for adiabatic case (Poisson's equation):

$$0 = c_p dT - \alpha dp \tag{11.2}$$

Using the equation of state and Clausius-Clapeyron equation, we get:

$$\frac{c_p}{RA}\frac{de_s}{e_s} = \frac{dp}{p} \tag{11.3}$$

where c_p is specific heat under constant pressure. By integrating Eq. 11.3 from the initial state $[e_s(T_0), p_0]$ to the final one $[e_s(T), p]$ for $\chi > 1$, we obtain:

$$\frac{E_s(T)}{e_s(T_0)} = \left(\frac{p}{p_0}\right)^{\chi} + \left(\frac{e}{e_0}\right)^{\chi} \tag{11.4}$$

where $\chi = AR/c_p \approx 5.7$ and e is vapour pressure. As the exponent $\chi > 1$ and $\frac{p}{p_0} < 1$, from Eq. 10.4, it is evidenced that our claim is true.

11.1.3 Water Vapour Supply by Air Mixing

When mixing any two saturated elements of air with different temperatures, the supersaturation appears in the mixture. That is because the water vapour density of the mixture is greater than the water vapour density of the saturated air corresponding to the temperature of the mixture. It is also possible that two unsaturated elements of air, whose temperatures are sufficiently different, by mixing become saturated. Let's observe two air particles found at the same atmospheric pressure p. With variables which characterize them (mass **m;** specific humidity **q;** water vapour pressure **e** and temperature **T**), we will denote with index (1) for the first individual air particle and (2) for the second. By mixing those air particles under constant pressure, we get a mixture whose features will be denoted without index. It is obvious:

$$q = \frac{m_1}{m_1 + m_2} q_1 + \frac{m_2}{m_1 + m_2} q_2 \tag{11.5}$$

In addition, with a great accuracy, it can be written as:

$$e \approx \frac{m_1}{m_1 + m_2} e_1 + \frac{m_2}{m_1 + m_2} e_2, \text{and} \tag{11.6}$$

$$T \approx \frac{m_1}{m_1 + m_2} T_1 + \frac{m_2}{m_1 + m_2} T_2 \tag{11.7}$$

This mixing could be understood if we look at the diagram shown in Fig. 11.2.

The curve $e_s(T)$ shown in Fig. 11.2 represents the Clausius-Clapeyron equation which gives relationship between the temperature of a liquid **T** and its saturated vapour pressure (e):

$$\frac{1}{e_s}\frac{de_s}{dT} = \frac{L_V}{R_V T^2} \tag{11.8}$$

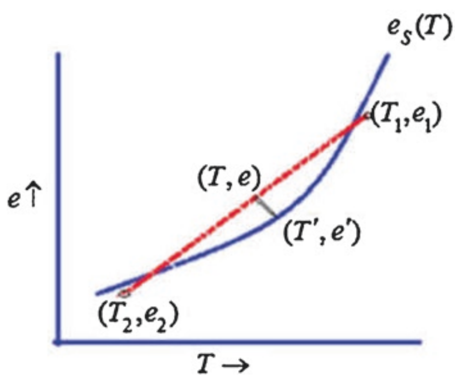
When the two air masses are mixed, the mixture can become supersaturated. The condensation process and the formation of the cloud occur. The condensation process will increase the temperature due to latent heat release ($-Ldq$). If this process is isobaric, then the rate of temperature increase is found by the relation:

$$\left(-Ldq\right) = c_p dT \tag{11.9}$$

or from the relation for **q** under given conditions:

$$dq = \frac{\varepsilon}{p} de \tag{11.10}$$

Fig. 11.2 Temperature as a function of vapour pressure



where $\varepsilon = 0.622$. By substitution of Eq. 10.10 into Eq. 10.9, we get the following expression:

$$\frac{de}{dT} = -\frac{c_p}{\varepsilon L}p \tag{11.11}$$

This formula determines the inclination length which describes the isobaric condensation process at the diagram (T, e). The cross-section of that length with curve $e_s(T)$ determines the point (T', e') that characterizes air mixture after termination of the condensation process. Mixing can also be observed under adiabatic conditions. If a two air parcels are found under different pressures, mixing will occur under equal pressures. The specific humidity will be described by Eq. 10.5, while the potential temperature is equal to the average potential temperature of each air parcel, analogously to Eq. 10.7. In such mixed column of air over unique surface, the specific humidity is given by:

$$q_m = \frac{1}{e_s}\int_{z_1}^{z_2} q\,dz \tag{11.12}$$

where m is the mass of the column of air.

11.1.4 Mixing and Diffusion

Mixing is always associated with bringing the water vapour into the observed air, or sometimes supersaturation can be achieved by water vapour supply in saturated environment through diffusion. Diffusion is important over large water surfaces when two air saturated masses meet and dense fog forms with light rain-drizzle and snow. Another way of supersaturation occurrence is when the moist air comes over water surface with different temperatures:

(a) When warm moist air comes over cold water surface, it is cooling from bellow. By cooling the air become supersaturated, but the stability of this layer of air parcel is growing. Fog that is formed under such conditions is shallow "sea-breeze fog".
(b) When cold air moves over warm surface, due to upward diffusion of water vapour, "evaporation or steam fog" forms.

11.1.5 Diabatic Cooling

A thermodynamic change of state of a system in which the system exchanges energy with its surroundings by virtue of a temperature difference between them. Temperature changes due to diabatic heating or cooling by them do not necessarily depend on whether the parcel is rising or sinking. Diabatic cooling mechanisms are:

- *Radiation* (heat loss by radiation occurs in area with large density gradient-nocturnal radiation cooling-radiation fog)
- *Molecular diffusion*-conduction (warm air moves over cold surface-formation of advection fog)
- *Turbulent diffusion*-convection (by turbulent diffusion and convection, heat is transferred vertically).

11.1.6 Formation of Cloud Elements

From the large number of processes leading to air saturation and formation of clouds in the troposphere, the most important is the adiabatic condensation in upward movements. Depending on the stability of the atmosphere and mechanisms that cause upward movement, the most diverse clouds appear, including cumulus, altocumulus, lenticularis, nimbostratus, or cirrus. When air rises it cools and the relative humidity increases. When a certain supersaturation is achieved in such environmental conditions, cloud droplets are formed on the soluble particles that serve as the cloud condensation nuclei CCN. The number of active particles (those on which drops are formed) increases with an increase in supersaturation. The number of formed droplets varies from $\left(\frac{30}{\text{cm}^3} \text{to} \frac{3000}{\text{cm}^3}\right)$. Such simply described process occurs at the cloud base, but it is more complex due to the mixing, falling of larger drops,

and accretion, which all tends to decrease supersaturation. By diffusion of water vapour, cloud drops grow rapidly at about 5 μm in diameter. The latent heat is released from drops by condensation. The growth of droplet is determined by equilibrium between both processes (diffusion and condensation). The growth of the droplet by diffusion is reversely proportional to the radius.

11.1.7 Precipitation

The term precipitation incorporates all hydrometeors (rain, snow, hail, dew, frost, rime, and other types). On the global scale, about 5% of all precipitation falls as snow as the result of high precipitation efficiency in low latitudes. Two mechanisms contribute in the formation of a significant number of large droplets from the small droplets, which actually form the precipitation (Rogers and Yau 1989; Ćurić 2001b):

1. Collision and collection-coalescence (warm rain)
2. Ice particles' growth into supercooled cloud water through ice nucleation (cold rain)

The second process occurs as a result of the existing difference between saturation vapour pressure over water and ice. That is suggested by Wegener (1911) and approved by Bergeron (1935), so-called Bergeron process. Langmuir (1948) examined the first process. In the real environment as cloud, both processes are competitive. As the cloud drops in the cloud are randomly distributed, it would not be possible to claim which one of the neighbouring drops will have the privilege to become larger by merging. A droplet that attains such privilege gets large fallout speed thus qualifying for more rapid merging. When the small droplet reaches the cloud radius, it continues to grow rapidly and forms raindrops with a radius of 1 mm. The diffusion growth of the droplet according to the law (1/r) indicates that the population of the droplets tend to be about of the same radius. This would diminish the possibility for collision and coalescence. However, in the population of the droplets whose number increases through diffusion, the mutual interaction among droplets rapidly increases. The fall speed of smaller droplets is proportional to r^2. From there, it shows that the fall speed affects the width of the spectrum of particle sizes. The higher the water content, the faster the intensity of the drops' interaction increases, as the efficiency of collection, the cross-section of droplet, and the difference in the falling rate also increase. Thus, the production of large droplets is highly dependent on the content of liquid water.

11.1.8 The Mechanism of Ice Nucleation

The mechanism of ice nucleation that leads to formation of precipitation (Heintzenberg and Charlson 2009) utilizes the condition that the maximum water vapour pressure over ice is lower than the vapour pressure over water at the temperatures lower than 0 °C. This mechanism is most pronounced at temperatures around −12°C when the difference between vapour pressures is about 0.27 hPa. Injection of ice crystals into the supercooled cloud, leads to ice crystal growth at the expense of water vapour and evaporation of cloud drops. Such cloud environment is unstable saturated as the result of the rapid change of water vapour content relative to warm clouds which represent stable saturated environment. When evaporating the droplets, the air is cooling, and because of the simultaneous deposition (transfer of water vapour to ice), the heat is released. The overall effect of these two processes leads to air heating.

11.1.9 Classification of the Microprocesses

Microprocesses of cloud formation and precipitation could be divided into three basic categories: nucleation phase, phase of growth or evaporation by water vapour diffusion, and phase of particle growth by accretion. Three separate phases appear at the nucleation process: nucleation of water from water vapour, nucleation of ice from water vapour, and ice nucleation from water. The equation which describes one category is similar to another, regardless of whether we consider water or ice. In any case it could be stated that a more simply task is to study the processes associate with water. These nucleation processes could be observed into clear environment (homogeneous nucleation) or into environment which contains additional substituents (heterogeneous nucleation). The growth and evaporation of particle by diffusion are being complicated due to the kinetic effects presented at small particles and also as the result of ventilation effects at larger particles and competition effects at particles found among a number of others. In intercellular interactions, four nucleation processes occur: 1) Coalescence (collection of the smaller particle by large particle); 2) Aggregation (collection of the smaller ice crystals by larger; 3) Freezing (collection of the smaller ice particle on large drop); 4) Accretion (collection of small water drops on large ice crystal). Here, larger, or particle that is fallout faster is called collector.

11.2 Cloud Definition and Classification

Clouds are among the most significant atmospheric phenomena. At any given time, they cover about half of the Earth's surface. Without clouds there would be no precipitation, thunders, lightning, and a wide variety of spectacular optical phenomena. They change the heat balance in the Earth-atmosphere system and cause different hydrometeors, electrometeors, and photometeors to occur in the atmosphere. Clouds with their shape and accompanying phenomena give a clear picture of the air currents at altitude and atmosphere.

Clouds represent a visible set of tiny water droplets and/or crystals of ice floating in the air. Clouds, like fog, are formed by the condensation and deposition of water vapour in the atmosphere, as well as the freezing of water droplets. Fog and cloud differ significantly, although they are similar in composition. Fog forms in the ground layer of the atmosphere during calm weather or in low winds, with clouds forming at different altitudes above ground, usually with strong vertical air currents. Water droplets and ice crystals of which the fog and clouds differ in their dimensions in that the clouds consist of slightly larger droplets and crystals than the fog. Fog is generally a stable formation, while clouds can be unstable, stable, and weak.

Clouds are in a continuous process of formation, development, and dissipation. At the border with cloudless air, cloudy droplets evaporate or ice crystals sublimate. As long as there is an influx of water vapour that condenses or becomes solid by deposition, the cloud does not disappear. Clouds disappear especially quickly when they are in a descending air stream, and because of the adiabatic warming, cloud droplets evaporate.

11.2.1 Cloud Classification

Clouds appear in a wide range of sizes and shapes. According to Wang (2013), the most suitable scientific classification of clouds is based on two main criteria (features):

1. Cloud visual patterns based on the observation
2. Cloud base height

This classification is initially proposed by the British meteorologist Luke Howard since 1803 and was later modified and adopted by the World Meteorological Organization (WMO). Even so, clouds can be classified into a number of groups based on their shape, height, method of formation, and composition.

11.2.2 Cloud Classification by Form

By form, clouds can be divided into three basic and multiple transitional forms. The three basic forms are cumulus, stratus, and cirrus. Cumulus, or bulky clouds, occur in an unstable atmosphere where vertical air movements are expressed. The rising air is local, so these clouds do not cover the entire sky, but have a cellular structure. Layered clouds, or strata, are formed by forced uplift or turbulent mixing of air in a stable atmosphere. Cirrus or feather-cloud clouds occur in areas of high-altitude winds.

11.2.3 Cloud Classification Based on Height

According to the height at which they occur, clouds are divided into low, medium, and high clouds. There is also a fourth group to which vertical development clouds belong. These clouds develop more vertically than horizontally, extending in a thicker layer of air. The height at which individual clouds from each of these groups are located depends on latitude. Table 11.1 gives the approximate values of cloud base height by geographical area. The differences in the altitudes of the cloud base are primarily the result of the fact that the altitude at which the air temperature is less than 0 °C changes with latitude. In the tropics, the zero isotherm is at a height of about 6 km and in the polar regions at about 3 km.

Table 11.1 Cloud base height (km) depending on latitude

Cloud levels	Tropical area	Mid-latitude area	Polar region
High	6–18	5–13	3–8
Middle	2-8	2-7	2-4
Low	Below 2 km altitude	Below 2 km altitude	Below 2 km altitude
Clouds with vertical development	Below 2 km altitude	Below 2 km altitude	Below 2 km altitude

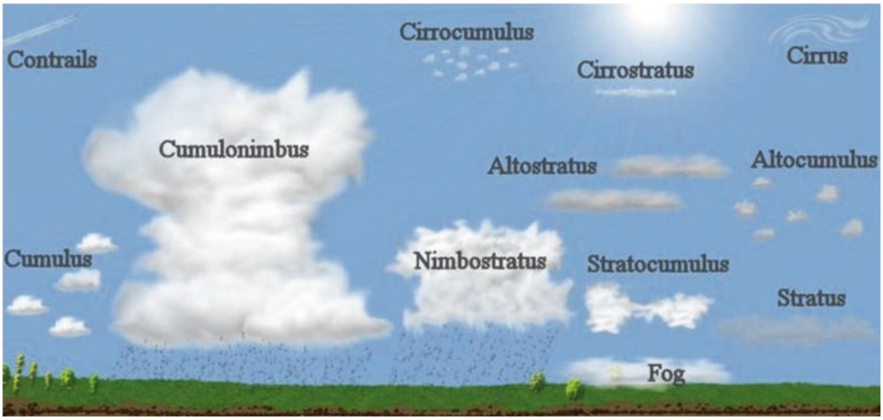


Fig. 11.3 Basic cloud types

11.2.4 International Cloud Classification

The international cloud classification, adopted by the World Meteorological Organization and described in detail in the Cloud Atlas, is made primarily by the shape and height at which the clouds are located (Fig. 11.3). Based on altitude, the clouds were classified into four groups and, based on the shape, into ten species (Table 11.2).

High clouds are almost exclusively composed of ice crystals due to the low air temperature at the altitudes at which they are formed. They are clouds of low thickness, mostly white in colour, from which no precipitation is extracted. Cirrus are high clouds that occur most frequently (Fig. 11.4).

These are separated clouds of fibrous or wavy appearance. Cirrocumulus is a high cloud that looks like clumps of snow, properly arranged in the form of folds or furrows. They usually cover a smaller portion of the sky. Cirrostratus are high clouds that cover the entire sky like a veil, through which the contours of the Sun or Moon are recognized. Very often, in their presence, a halo can be observed. These clouds indicate the arrival of the warm front. Cirrus clouds have five defined subclasses: fibratus, uncinus, spissatus, floccus, and castellanus.

Table 11.2 International cloud classification

Cloud levels	Type
Low	Nimbostratus (Ns) Stratocumulus (Sc) Stratus (St)
Middle	Altocumulus (Ac) Altostratus (As)
High	Cirrus (Ci) Cirrostratus (Cs) Cirrocumulus (Cc)
Clouds with vertical development	Cumulus (Cu) Cumulonimbus (Cb)

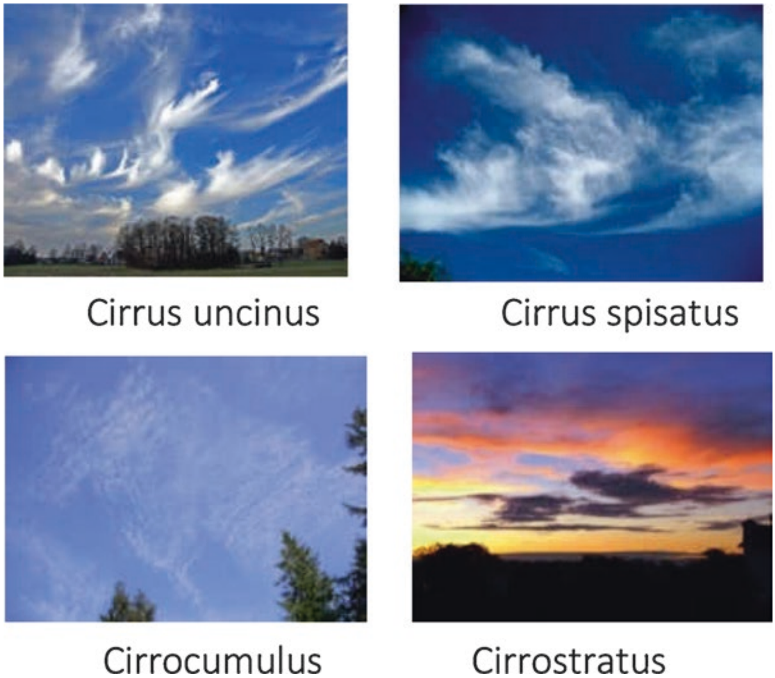


Fig. 11.4 High clouds

Middle clouds are usually made up of cold water droplets, but they can also be of mixed composition or even composed of ice crystals only. Altocumulus is a grey or white cloud in the form of clumps or plaques, usually properly spaced (Fig. 11.5). They differ from the circus cumulus, except in height and in that the elements of the cloud layer are larger, and usually their inner part is lighter, and the exterior is darker in colour. Rainfall may occur from the altocumulus. Altostratus (altostratus) is a grey or bluish, furrowed, or uniform cloud layer that usually covers the entire sky. Through the thinner parts of this cloud, the Sun can be seen as through the milk

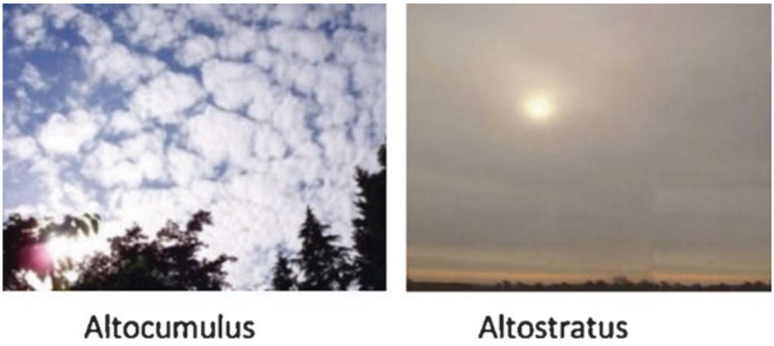


Fig. 11.5 Middle clouds

Fig. 11.6 Low clouds

glass. They are different from cirrostratus in height, colour, and the way they transmit sunlight. They do not cause shadow on the Earth and do not cause halo. Low-intensity rainfall over a wider area can be extracted from the altostratus.

Low clouds are mostly made up of water droplets. At low temperatures, they may also contain snow and ice crystals. Nimbostratus (nimbostratus) is a dark grey cloud of considerable thickness, completely obscured by the Sun and Moon. It is a precipitation cloud from which rain or snow from continuous to moderate intensity is continuously falling (Fig. 11.6). Visibility below the nimbostratus is often very diminishing as the falling rain evaporates. If water vapour is saturated below the nimbostratus, fog or even lower cloud formation may occur.

A stratocumulus (stratocumulus) is a grey-white layer composed of separated or assembled plates, pebbles, or rollers. The elements of this cloud layer are larger than those of the altocumulus. From this cloud, occasional rain or snow falls. Stratus is a low grey cloud covering the entire sky. Its composition is like fog but does not touch the Earth's surface. Sometimes light rainfall, such as drizzle or light snow, can fall from it.

Vertical development clouds are generated by intense vertical uplift of air in a statically unstable atmosphere (Ćurić 2001a). Cumulus (cumulus) are isolated, clearly fringed clouds (Fig. 11.7). The base is flat and darker in colour, and the upper part is white and in the form of domes, hills. The cumulus of lower vertical

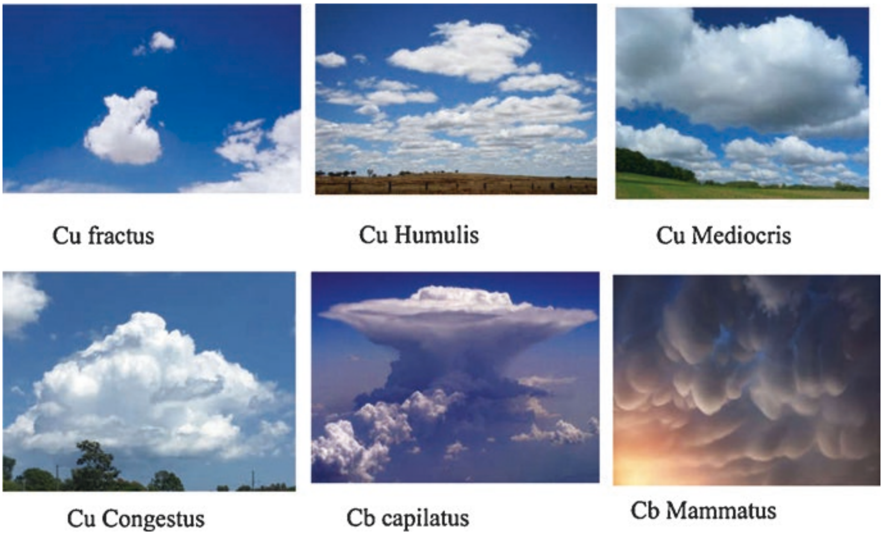


Fig. 11.7 Clouds with vertical development

development is the so-called nice weather cumulus, while larger and more developed cumulus that resembles cauliflower may produce drizzle.

Cumulonimbus clouds are clouds with vertical development looking as multilevel clouds, extending high into the sky in towers or plumes. More commonly known as thunderclouds, cumulonimbus is the only cloud type that can produce hail, thunder, and lighting. The base of the cloud is usually flat with a very dark base and may only lie a few hundred metres above the Earth's surface. Towering cumulonimbus extend from low altitude to very high altitudes. These are "thunderstorm" clouds and may produce lightning, thunder, and hail. This is the view of the bottom portion of a cumulonimbus cloud when it is about to rain. Because of its massive size and towering height, this is often the only view of a cumulonimbus that can be seen at close range. These often appear on the underside of a cumulonimbus and indicate the possibility of severe weather. Cumulonimbus clouds are born through convection, often growing from small cumulus clouds over a hot surface. They can form along cold fronts as a result of forced convection, where milder air is forced to rise over the incoming cold air. Cumulonimbus clouds are associated with extreme weather such as heavy torrential downpours, hailstorms, lightning, and even tornadoes. Individual cumulonimbus cells will usually dissipate within an hour once showers start falling, making for short-lived, heavy rain. However, multicell or supercell storms contain many cumulonimbus clouds, and the intense rainfall may last much longer. If there is thunder, lightning, or hail, the cloud is a cumulonimbus, rather than nimbostratus. Cumulonimbus clouds have a three distinct "species" which describe the appearance of the head of the cloud:

- **Cumulonimbus calvus** – The top of the cumulonimbus is puffy, like a cumulus cloud. The water droplets at the top of the cloud tower have not frozen to become ice crystals.
- **Cumulonimbus capillatus** – The top of the cloud is fibrous but relatively contained. Water droplets have started to freeze, usually indicating rain has begun or will begin soon.
- **Cumulonimbus incus** – The top of the cloud is fibrous and anvil shaped as the cloud has continued to grow. If the cloud reaches the top of the troposphere and still wishes to grow, it must do so outwards, creating the pictures que anvil or "incus".
- **Cumulonimbus mammatus** – Is a specific cellular pattern of hanging protuberances, like pouches on the under surface of a cumulonimbus cloud base, although they may be coupled to other classes of main clouds. They are formed by cold air sinking to form a small patch opposite to the puffs of clouds rising through the convection of warm air.

In addition to 10 general cloud types, there are 14 species, 9 subspecies, and 9 special features that distinguish clouds. A cloud of a type can belong to only one species, while a certain species can be attributed to clouds of different genera. The cloud may have characteristics of multiple subspecies, meaning that subspecies are not mutually exclusive as species. The same is true of additional cloud features, i.e. the viewed cloud can have one or more complementary features. The name of a species, subspecies, or additional feature may not necessarily be attributed to a cloud unless it is clearly expressed. Only some species, subspecies, and complementary features will be mentioned here: lenticularis (lenticular; species Cc, Ac, or Sc), mamma (breast; additional marking Ci, Ac, As, or Sc with a hanging pouch), fructus (broken; type St or Cu), humilis (small size; type of poorly developed Cu usually flat base), mediocris (type of cumulus of moderate vertical development, the upper parts of which show moderately developed protuberances), congestus (accumulation; type of highly developed Cu in the vertical direction, cauliflower-like), calvus (bald; type Cb losing its bulky form usually at the top), capillatus (hairy; type Cb fibrous upper), incus (anvil, additional marking Cb with anvil-shaped upper), pileus (cap; additional feature of cumulus caps with hooded cap, hood), castellanus (castle; a type of tower-like vertical development cloud), undulatus (wavy, wavy-shaped cloud subspecies), translucidus ((translucent; cloud-covered subspecies), i.e. most of the sky and through which the sun or moon is seen), (precipitation; an additional feature of As, Ns, Sc, St, Cu, or Cb from which precipitation is extracted)), etc.

11.2.5 Cloud Classification by Composition

The clouds may be composed of water droplets, ice crystals, or mixed composition. The dimensions of water droplets, as well as the dimensions and shape of ice crystals, can be very different and in the same cloud. Depending on the composition, clouds may be:

1. Stable if composed of ice crystals or tiny water droplets
2. Unstable if mixed or consisting of larger water droplets of different dimensions
3. Weakly volatile when composed of ice crystals or smaller water droplets of different dimensions

Unstable clouds are mostly rainfall clouds, while there is generally no precipitation from stable clouds. It can occasionally pour rain or drizzle from low clouds. The most stable are the high clouds (cirrus, cirrostratus, cirrocumulus) and the most unstable are cumulonimbus.

11.2.6 Classification by Mechanism of Formation

Clouds are most commonly generated by adiabatic expansion and cooling of air due to thermal convection, rising air at orographic obstacles, rising warm air above the wedge of cold air, dynamically conditioned vertical air movements, or a sudden local pressure drop that occurs in tornadoes, thrombi, and seas (Ćurić 2001b). Cloud formation can also occur due to the turbulent transport of water vapour and condensation nuclei to the inversion layer and the mixing of air masses at different temperatures, as well as wave motions in the atmosphere.

Orographic clouds are generated by the forced uplift of moist air when approaching a mountain barrier (Fig. 11.8). They can occur in front of, above, or

Fig. 11.8 A rotor cloud as type of orographic clouds

behind an orographic obstacle and take many forms. If the atmosphere is unstable, cumulus clouds are formed, and layer clouds are stable.

When transferring air through mountain ranges, on the windward side, large-scale clouds form, which cover the mountain top as a cap, and are called a hair dryer or a wall of hair. They can also be seen on the windward side of the mountain, and if the air is very humid, they can also cross the mountain ridge to the height at which all cloud droplets evaporate when descending. Under certain meteorological conditions, the formation of so-called standing waves when transferring air over mountain ranges occurs. On the ascending parts of standing waves, if the air is moist enough, clouds of lenticular forms of the genus Sc, Ac, and Cc, of the lenticular species, occur. Downstream currents in the lower layers of the atmosphere are vortices with a horizontal axis, the so-called rotors, and if the air is sufficiently damp on the upstream part of the vortex, roll-shaped thick clouds. In addition to the described orography wave, air waves can also occur in the free atmosphere under certain meteorological conditions (e.g. horizontal movement at unequal speeds of two air masses at different temperatures, below altitudinal inversions, etc.). If the air is humid enough, condensation and deposition of water vapour and the formation of wave clouds, most commonly of the genera As and Sc, occur.

11.2.7 Special Clouds

In the atmosphere, in addition to the clouds described so far, which form in the troposphere, clouds form in the higher layers of the atmosphere. In the stratosphere, mother-of-pearl clouds, a luminous iridescent cloud, can be viewed (Fig. 11.9) and night-light clouds in the mesosphere. Mother-of-pearl clouds occur at altitudes between 20 and 30 km. They can be viewed during the winter months in the mountainous polar regions (Alaska, Scandinavia, etc.). In sunlight, they have a brilliant

Fig. 11.9 A luminous iridescent cloud

mother-of-pearl colour. On the mother-of-pearl clouds, there is also irritation, so colours that appear banded and almost parallel to the edge of the cloud can be seen. It is assumed that mother-of-pearl clouds result from adiabatic cooling at the crest of orographic waves reaching higher layers of the atmosphere. The night luminescent clouds of the upper mesosphere, consisting of small crystals of ice formed on particles of meteoric origin, are described in more detail in the temperature section.

11.3 Fog

Fog is defined as a visible aerosol consisting of tiny water droplets or ice crystals suspended in the air. Fog droplets are found with different dimensions. In supercooled temperatures, they reach 2–5 microns, while in the positive temperature, 7–15 microns. When the visibility of air is reduced below 1000 metres, fog occurs. Fog is a cloud base at the Earth's surface or very close to it, and it is considered as an unfavourable temporal phenomenon.

11.3.1 Fog Types and Formation

To form fog, the air must be saturated with water vapour. The relative humidity in conditions of fog is around 100%. Fog is caused by tiny water droplets suspended in the air. The thickest fogs tend to occur in industrial areas where there are more pollution particles in the air allowing water droplets to coalesce and grow. Additional favourable atmospheric conditions for its occurrence are: high air pressure (anticyclone weather), stable atmosphere and appearance of the temperature inversion, no wind or very weak flow of air, suitable topographic conditions, and the presence of condensation nuclei. In Fig. 11.10, different forms of fog which is formed by cooling are illustrated:

1. Radiation fog (formed by radiation cooling of the Earth and the adjacent air)
2. Advective fog (generated when warm and moist air flows over a cold surface)

Fig. 11.10 Radiation fog

Fig. 11.11 Evaporation fog

Fig. 11.12 Urban and industrial fogs

3. Orographic fog (occurs when relatively damp air rises up at the topography and cools adiabatically)

Fogs which are formed through evaporation (Fig. 11.11) are divided into:

- Evaporation fog (when water vapour that rises over warm water condenses into cold air)
- Frontal fog (when warm air rises over cooler air along the front)

For fogs forming in urban areas, despite the above-mentioned conditions for formation, there are additional factors (stagnant air, higher concentrations of polluting substances) that stimulate the process of fog formation. These are known as urban fogs which usually occur in combination with smog. Fogs that occur in industrial zones in surface mines, lignite coal, and other mining and energy sites are called industrial fogs (Fig. 11.12).

These fogs are a combination of pollution at the ground layer of atmosphere and moisture, which are very favourable conditions for the formation of dense fog, unable to reduce to just a few metres horizontal and vertical visibility in the atmosphere.

Fig. 11.13 Freezing fog

Supercooled fog. Supercooled fog forms in similar way as radiation fog due to a nigh-time cooling of the Earth's surface at clear sky conditions when heat is rapidly transferred back to the space. During this process, water vapour condenses into tiny water droplets that exist at temperatures colder than 0°C, i.e. they become supercooled water droplets remaining liquid even though they are below freezing temperature (Fig. 11.13). When supercooled water liquid droplets freeze onto surfaces, a white deposit of feathery ice crystals is formed.

11.4 Hydrometeors

Any water or ice particles that have formed in the atmosphere or at the Earth's surface as a result of condensation or sublimation. Water or ice particles blown by the wind from the Earth's surface are also classed as hydrometeors. According to the AMS Glossary, the following classification of hydrometeor is adopted:

1. Liquid or solid water particles formed and remaining suspended in the air, for example, damp (high relative humidity) haze, cloud, fog, ice fog, and mist
2. Liquid precipitation, for example, drizzle and rain
3. Freezing precipitation, for example, freezing drizzle and freezing rain
4. Solid (frozen) precipitation, for example, snow, hail, ice pellets, snow pellets (soft hail, graupel), snow grains, and ice crystals
5. Falling particles that evaporate before reaching the ground, for example, virga
6. Liquid or solid water particles lifted by the wind from the Earth's surface, for example, drifting snow, blowing snow, and blowing spray

Most frequent forms of precipitation at the Earth's surface are:

- Rain (water drops that fall from the clouds and have a diameter smaller than 0.5 mm)
- Snow (precipitate in the form of ice crystals or, often, a set of ice crystals)

11.4.1 Precipitation

The rain is a precipitate which is formed when separate drops of water fall on the Earth's surface from the clouds. Not all the amounts of rain reach the soil. Some of the raindrops evaporate while falling through dry and warmer air. When the raindrops do not reach the Earth's surface, the phenomenon is called virga, and it is often controlled by the hot and the dry regions. The scientific explanation of how rain is formed and how it falls to the surface is called Bergeron process. Rain plays a role in the hydrological cycle in which moisture from the oceans, seas, lakes, and the land evaporates, condenses into droplets that form clouds, falls to the Earth's surface in the form of rain, and again returns to the air in order to continue the round cycle. Rain, as an atmospheric phenomenon, is characterized by two categories: the quantity of rainfall and the factor for the occurrence of rainfall. Based on the amount of precipitation in the observed time interval, the rainfall intensity could be classified into the following categories:

- Very weak rain – the intensity rainfall is <0.25 mm/hour.
- Weak rain – the intensity of rainfall is 0.25–1.0 mm/hour.
- Moderate rain – the intensity of rainfall is 1.0–4.0 mm/hour.
- Heavy – the intensity of rainfall is 4.0-16.0 mm/hour.
- Heavy rain – the intensity of rainfall is 16.0–50.0 mm/hour.
- Extreme rain – the intensity of rainfall is > 50.0 mm/hour.

Figure 11.14 shows the different forms of precipitation that occur in the atmosphere: hail, snow, and rain. According to the mechanism of occurrence, precipitation is classified into:

- Orographic rain
- Convective rain
- Frontal/cyclonic rain

Fig. 11.14 Form of rainfall (hail, snow and rain)

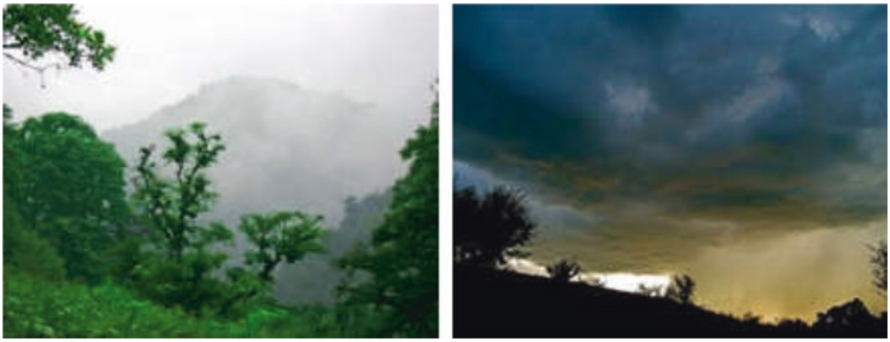


Fig. 11.15 Orographic rainfall

11.4.2 Orographic Rainfall

Orographic rain occurs during warm and wet wind blowing from the sea to the mainland, which encounters the natural barrier created by the mountains (Fig. 11.15). This obstacle forces the air to rise. As the altitude increases, due to the drop in air pressure, the air expands dynamically and reduces the air temperature, which results in an increase in relative humidity. This causes condensation of water vapour into water droplets and formation of clouds. The relative humidity continues to rise until the dew point reaches the level of condensation, causing air saturation. The altitude where condensation occurs is called level of condensation (CCN). When the clouds become heavy enough to hold the condensed cloud mass, it starts raining. When the wind is blowing downslope of the leeward side of the mountain, it is compressed and it warms up, which results in further reduction of the relative humidity of the wind that is already dry after rainfall, manifested in the upwind side of the mountain. Hence, at the leeward side of the mountain, rainfall from these winds does not appear.

11.4.3 Convective Rainfall

Convective rainfall mainly occurs in equatorial and tropical climatic regions, where the days are very hot and humid. This type of rainfall usually occurs in mid-latitude regions, more frequently in the warm part of the year. The speed of evaporation of moisture from water bodies and respiration of thick vegetation is very high. Evaporated moisture and the surrounding warm air begin to rise. By achieving the height, the air expands dynamically, and this phenomenon is due to the reduction of air pressure. As a result, a decrease of temperature occurs, which results in an increase of relative humidity, which causes the condensation of water vapour into water droplets. In this form, unstable clouds with vertical development (cumulonimbus) provide intense rainfall showers (Fig. 11.16). Convective rainfall is usually

Fig. 11.16 Convective rainfall

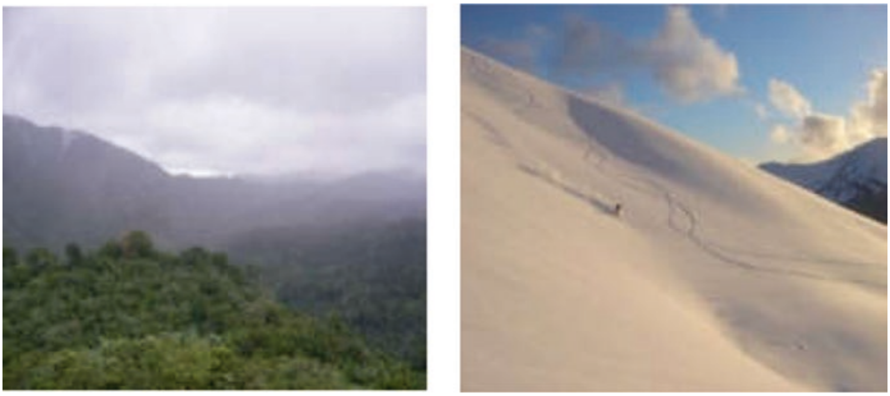


Fig. 11.17 Frontal rainfall and snowfall

occurring late at noon, and it is accompanied by atmospheric electrical discharges, lightning, and thunder. Convective rain is usually associated with tropical depression (typhoons) and storms.

11.4.4 Frontal Rainfall

Frontal or cyclone rainfall is caused by cyclonic activity and appears along the front of the cyclone. Depending on the characteristics of air masses, geographic latitude, and atmospheric conditions, such as front and physical processes, the frontal rainfall may be rain, snow, or hail (Fig. 11.17). In general, precipitation is formed when two air masses with different temperature, humidity, and air density encounter. The layer that separates these air masses is called a front. Frontal precipitation usually occurs when a warm front encounters a cold front. The less dense warm air parcel rises and cools into the ascedental flow in region of strong updrafts. The heavier cold sinks at the near surface and form clouds. At the frontal boundary, usually the clouds bring heavy precipitation.

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Chapter 12
Atmospheric Motion

The air in the atmosphere is in continuous motion, as the result of non-equal distribution of energy over the Earth's surface. Different energy distribution, which is manifested by the different heating of the Earth's surface, creates differences in pressure. The pressure differences produce a horizontal air motion which is known as wind. Atmospheric movement is caused by forces that affect horizontal and vertical movement of air. Dynamic meteorology deals with movements that significantly determine weather and climate (Morel 1973; Ćurić 1983, 2000, 2002 Gill 1982; Holton 2004; Mak 2011; Lin 2007; Markowski and Richardson 2010; Saha 2008; Spiridonov and Ćurić 2011; Ćurić and Janc 2016; Zdunkowski and Bott 2003; Vallis 2017). In this chapter only the basic elements of dynamics of the atmosphere are described. In all these cases, the atmosphere does not need to be studied through the individual molecules that it is composing but can be considered as a continuous fluid medium or briefly a continuum (Hoskins and James 2014). Any point in that continuum will in fact be considered as an elementary volume that is very small compared to the volume of the part of the atmosphere that is subject to interest, but it still contains a large number of molecules. We will always use the term "air particle" when we refer to the characteristics of a certain point of space. The state of the atmosphere is characterized by various meteorological variables (e.g. pressure, temperature, density, etc.). It is assumed that each point of the atmosphere corresponds to one (unique) quantity of these variables. It is assumed further that these variables and their derivatives are continuous functions of space and time.

The basic laws of fluid mechanics and thermodynamics that determine the movement of the atmosphere can be described by differential equations. Here the fluid characteristics are considered as dependent variables, while time and space (location) are independent variable quantities. The atmospheric motion (Fig. 12.1) follows a three fundamental conservation laws: the law of conservation of mass, momentum, and energy (Lindzen 2005; Spiridonov and Ćurić 2011; Zdunkowski and Bott 2003). These principles are applied on a small elementary volume of the atmosphere to derive the corresponding basic equations of motion.

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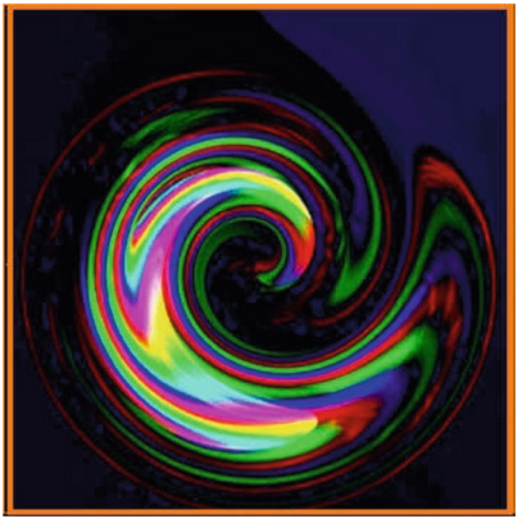


Fig. 12.1 Graphical illustration of fluid motion

12.1 Real Forces

Before to derive a complete set of equations which describes the atmospheric motion, it is useful to explain the nature of the forces that cause the movement of the atmosphere. It is known that various forces act on the air movement into atmosphere. The basic and most important atmospheric forces are those which act independently of whether the Earth would rotate or be in a state of rest. They would be the same in this rotating atmosphere and in the same situation when the Earth would not rotate. From the point of view of the *Newton's second law of motion*, it can be said that they are forces that act in the action that is observed in relation to the coordinate system whose axes are fixed in the space. The most important forces regarding to the atmospheric motion are pressure gradient force, gravitational force, and friction force. These forces are called *external* or *real* forces.

12.1.1 Pressure Gradient Force

There is no uniform distribution of pressure in the surrounding atmosphere. To explain the pressure gradient force, it is necessary to describe the meaning of pressure gradient. In atmospheric science (meteorology and the related disciplines), *the pressure gradient* (typically of air, more generally of any fluid) is a physical quantity that describes in which direction and at what rate the pressure increases the most rapidly around a particular location. The pressure gradient is a dimensional quantity expressed in units of pressure per unit length or Pa/m. The pressure gradient force results in a net force that is directed from high to low pressure, and this force is called the "pressure gradient force". It is responsible for triggering the initial movement of air in the atmosphere. Horizontal variations of air pressure are much weaker than vertical pressure variations, only about one ten-thousandth as large. But there's no horizontal gravity force to counterbalance them. Thus, the horizontal pressure gradient force generally forces the air to move, producing wind. Horizontal pressure gradients work just like vertical pressure gradients: a stronger pressure on one side of an air parcel than the other causes a net force in a direction. The pressure gradient force acts on any individual air particle in the atmosphere. Let this element, bounded by the surface S, have a volume of V. On the elementary surface dS, we can take that the pressure p is equal at each point. It follows that the pressure force p = − p dS **n**, Fig. 12.2, where **n** is an external normal to an element of surface dS. The pressure force component in the direction of the x-axis is $p_s = -p\,dS\,\mathbf{n} \bullet \mathbf{i}$.

The resulting component of all forces which act on the surface S in the direction of the x-axis is:

$$G_{mFX} = -\int p\mathbf{n}\cdot\mathbf{i}dS. \tag{12.1}$$

Hence, using Gaussian identity, it is obtained:

$$G_{mFX} = -\int \nabla\cdot(p\mathbf{i})dV = -\int \mathbf{i}\cdot\nabla p dV, \tag{12.2}$$

since $\nabla \bullet \mathbf{i} = \mathbf{0}$.

If the air element is small, we can assume that $\mathbf{i} \bullet \nabla p$ is independent which volume dV is considered, so the expression for the component of the pressure gradient force in x-direction is obtained:

$$G_{mFx} = -V\mathrm{i}\cdot\nabla p = -V\frac{\partial p}{\partial x}. \tag{12.3}$$

or using $V = m/\rho$, we get the pressure gradient force in x-direction per unit mass:

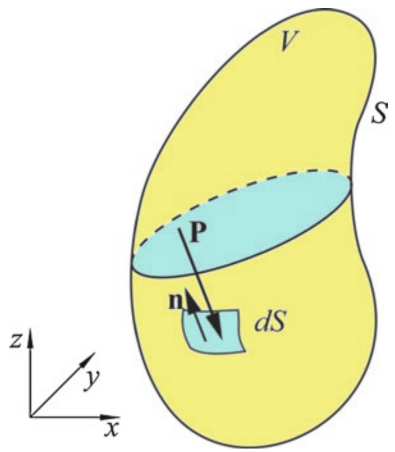


Fig. 12.2 Pressure gradient force acts on the elementary surface

$$G_{rx} = \frac{\mathrm{G}_{\mathrm{mFx}}}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial x} \tag{12.4}$$

By applying the same procedure, components of the pressure gradient force in y- and z-axis are obtained:

$$G_{ry} = \frac{\mathrm{G}_{\mathrm{mFy}}}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial y}; G_{rz} = \frac{\mathrm{G}_{\mathrm{mFz}}}{m} = -\frac{1}{\rho}\frac{\partial p}{\partial z} \tag{12.5}$$

In vector form, the pressure gradient force is given by:

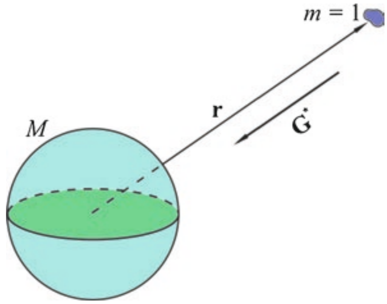
$$G_r = -\frac{1}{\rho}\nabla p. \tag{12.6}$$

The magnitude of the pressure gradient force is directly proportional to the size of the gradient of pressure, but its direction is the opposite. To recall the transfer of energy, the method of conductivity is applied. In heat conductivity, heat is transferred from an area with higher temperature to an area with lower temperature. The strength of the pressure gradient behaves in the same manner, i.e. it directs the areas of higher pressure to areas of lower pressure. This contributes to the air flow from regions with high pressure to regions with lower pressure.

12.1.2 Gravity

Atmosphere is found in field of Earth's gravitation, which acts on each individual air particle. Gravity is a force which is generated by the Earth and which attracts all objects towards its centre (Fig. 12.3). To assume that the atmosphere is a sphere is not a big mistake, since the difference between the largest radius of the Earth (6378 km in Equator) and the smallest (6357 km in the field) is only 21 km. Thus, an expression for the force which the Earth attracts an individual air particle with unit mass is in the form of:

Fig. 12.3
Gravitational force



$$G^* = -\gamma \frac{M}{|\mathbf{r}|^2}\left(\frac{r}{|\mathbf{r}|}\right) \tag{12.7}$$

where **r** is the vector directed from the Earth's centre towards the centre of individual air parcel, $\gamma = 6.66 \bullet 10^{11} m^3 kg^{-1} s^{-2}$ the gravitational constant, and $M = 5.988 \bullet 10^{24}$ kg the mass of Earth. It is evident that **G** depends on the distance between the centres of mass of the Earth and air of this element $r \equiv |r|$. Therefore, we can define the gravitation potential field:

$$\Phi' = -\gamma M\left(\frac{1}{r} - \frac{1}{r_p}\right) \tag{12.8}$$

where r_p is arbitrarily chosen distance from the centre of the Earth. Thus, G^* can be expressed as follows:

$$G^* = -\nabla\Phi' \tag{12.9}$$

If the distance of the air parcel from the centre of the Earth is expressed in terms of the mean Earth's radius R and height z above mean sea level (MSL), the gravity force gets the following expression:

$$G^* = -\gamma \frac{M}{(R+Z)^2}\left(\frac{r}{|\mathbf{r}|}\right) \tag{12.10}$$

Gravity force which acts on the air parcel with a unit mass m=1, found MSL, is:

$$G^* = -\gamma \frac{M}{R^2}\left(\frac{r}{|\mathbf{r}|}\right) \tag{12.11}$$

From here and from (12.10), we obtain:

$$G^* = \frac{G_n^*}{\left(1 + \frac{Z}{R}\right)^2} \tag{12.12}$$

We see that the gravitational force at any level can be identified with the gravitational force which acts in the same air parcel, in case it is located at a distance R from the centre of the Earth. This follows from the fact that the air parcels which are significant for meteorology are found at altitudes up to 20 km. This means that in this case R>>Z, it follows that $G_n^* = G^*$. Gravity force which acts on individual air parcel is called gravity acceleration g^*.Taking the expression (12.11), using $R = 6.333 \bullet 10^6$, the approximate rate of gravity acceleration is $g^* = 9.84\ ms^{-2}$.

12.1.3 Friction

There are various air movements in the atmosphere. There is a movement of molecules, which in general is chaotic, the movement of elements of air (whose mass is very different), and there is a movement of whole layers of air. Hence, it could be conditionally assumed that the atmosphere represents a set of molecules, a set of elements of air and air layers. All these “constituent” elements that are numerous have some characteristics: speed, amount of movement, temperature, etc. Movement takes place continuously transferring these characteristics from the place to the place to the other. This is the result of the most diverse changes in the quantity and quality of the atmosphere at any point. Here we are interested in a force that acts on any element of air and is the consequence of all mentioned types of motion. We call this force a force of friction or a viscous force. Let’s observe the element of air volume δx, δy, δz, found in the field of horizontal flow $\boldsymbol{v} = (u, v)$, which changes only with height (Fig. 12.4). Such a flow is valid to observe because in the atmosphere, wind changes with height are higher than in horizontal direction. Due to the above-mentioned air movements, the vertical transport of the momentum will be exerted.

In the absence of friction of the air, an individual air particle (object) would underline a constant acceleration in the vertical direction. The reason for this acceleration is gravity. Friction works against the movement. If an air particle moves up (rises), the friction is directed downwards and vice versa. Particles can reach the level where the influence of friction is approximately zero. This happens in the free atmosphere above the friction layer. In the layer below the free atmosphere, friction plays a major role in air movement. In a situation like Fig. (12.4), the friction force in the direction of the x-axis will act on the upper surface of the observed air element (τ_{zx}). Assuming that this force is proportional to the change of the velocity

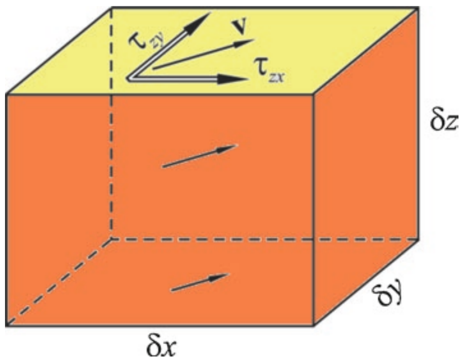


Fig. 12.4 Friction force which acts on the elementary mass of air whose volume is δx, δy, δz

component (u) with height, the expression for this force of motion is obtained or as it is called (shearing stress due to the wind shear) per unit area:

$$\tau_{zx} = -\mu \frac{\partial u}{\partial z} \tag{12.13}$$

where $\mu = \mu(z)$ is friction coefficient (has the same meaning as the dynamic viscosity coefficient but with much larger quantity). On the lower surface, the force per unit area will act in the direction opposite to x-axis $-\left(\tau_{zx} - \frac{\partial \tau_{zx}}{\partial z}\delta z\right)$. Resulting force in the direction of x-axis:

$$F_{xm} = \tau_{zx} - \left(\tau_{zx} - \frac{\partial \tau_{zx}}{\partial z}\delta z\right) = \frac{\partial \tau_{zx}}{\partial z}\delta z \tag{12.14}$$

As the mass of the air $\rho\delta z$ is found above the unit surface in the observed element, this can be defined by the friction force per unit mass:

$$F_x = \frac{1}{\rho}\frac{\partial \tau_{zx}}{\partial z}. \tag{12.15}$$

Similarly, it is obtained for the force of friction in the direction of the y-axis:

$$F_y = \frac{1}{\rho}\frac{\partial \tau_{zy}}{\partial z}. \tag{12.16}$$

Under the above assumptions, on the element of air will act the force which has the components:

$$\boldsymbol{F} = \left(\frac{1}{\rho}\frac{\partial \tau_{zx}}{\partial z}, \frac{1}{\rho}\frac{\partial \tau_{zy}}{\partial z}, 0\right) \tag{12.17}$$

It can easily be determined, if the observed element of air is in the flow field $\boldsymbol{\vartheta} = \boldsymbol{\vartheta}(x, y, z)$ where $\boldsymbol{\vartheta} = (\mathbf{u}, \mathbf{v}, \mathbf{w})$.The stress (force per unit area) at a point in a fluid needs nine components to be completely specified, since each component of the stress must be defined not only by the direction in which it acts but also the orientation of the surface upon which it is acting. The set of nine scalar variables is referred to us as stress tensor. It should be noted that the viscosity coefficient μ thus defined has a higher value than in laboratory conditions. In the atmosphere μ changes with height and in the horizontal direction. With the same wind shear (e.g. $\frac{\partial u}{\partial z} > 0$), the frictional force, i.e. μ, will be greater in case of small-scale processes, as they increase the mixing (transport) of momentum. Also μ is larger in areas with a strong convection than to regions with stable atmosphere. For the same reason, it is larger when the air moves across the irregular relative to the flat surface.

12.2 The Forces That Are the Effects of Earth's Rotation

It is most appropriate that air movement in the atmosphere is observed in relation to the coordinate system that is stationary in relation to the Earth. In relation to such a coordinate system, in practice, the wind speed is determined. But the introduction of such a coordinate system requires the introduction of additional forces (in addition to the stated – external forces). Namely, the element of air, which is in a state of rest in relation to the coordinate system related to the Earth, is neither in rest nor in a state of uniform motion in relation to the coordinate system fixed in the space inertial system. Newton's law of motion can be applied for motion in such a moving coordinate system only when considering the acceleration of that system in relation to the fixed system. Considering the acceleration of the coordinate system related to the Earth in relation to the fixed system in the space requires the introduction of "apparent" forces, centrifugal and Coriolis.

12.2.1 Centrifugal Force

Let us consider point A which rotates with uniform angular velocity Ω around some axis, Fig. 12.5. It can be any point that is fixed to Earth. In that case Ω is angular velocity of the Earth. Peripheral velocity of point A is:

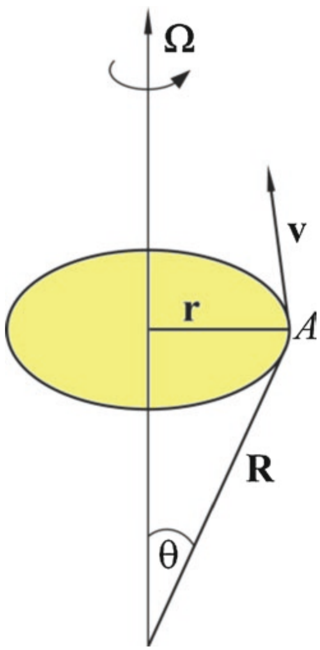


Fig. 12.5 Periphery velocity of point which rotates with constant angle velocity

$$\mathbf{v} = \mathbf{\Omega x R}, \text{or} \tag{12.18}$$

$$\left|\mathrm{v}\right|^2 = \Omega^2 \mathrm{R}^2 \sin^2\theta, r = R sin\theta. \tag{12.19}$$

Radial acceleration is:

$$\mathbf{a_r} = -\frac{\mathrm{v}^2}{\mathrm{r}}\frac{\mathbf{r}}{\mathrm{r}} = -\Omega^2 \mathrm{R}\sin\theta\frac{\mathbf{r}}{\mathrm{r}}. \tag{12.20}$$

The previous expression can be written in vector form:

$$\mathbf{a_r} = -\mathbf{\Omega x}\left(\mathbf{\Omega x R}\right). \tag{12.20}$$

The last term defines a centrifugal acceleration. This is also the term for the centrifugal force per unit mass. We'll denote it with **Z**. We see the centrifugal force with the function of the distance from the axis of rotation. It can be written (using 12.20) in the form:

$$\mathbf{Z} = \Omega^2 \mathrm{r}\nabla\mathrm{r} / \left|\nabla\mathrm{r}\right| \mathrm{if} \left|\nabla r\right| = 1, \mathrm{i.e.}\mathbf{Z} = \nabla\left(\frac{1}{2\Omega^2\mathrm{r}^2} + \mathrm{const}\right). \tag{12.21}$$

We notice that we can introduce the potential:

$$\Phi'' = -\frac{1}{2}\Omega^2 r^2 + \mathrm{const} \tag{12.22}$$

whose gradient is the centrifugal force. Thus, it follows:

$$\mathbf{Z} = -\nabla\Phi'' \tag{12.23}$$

The radial acceleration defined by the expression (12.20) is the largest on equator. For example, the Earth turns around its axis for 23 h 56 min and 4 s (i.e. 86164 s), so that:

$$\Omega = \frac{2\pi}{24 \cdot 60^2 - 4 \cdot 60} s^{-1} = 7.292 \cdot 10^{-5} s^{-1}. \tag{12.24}$$

For r = 6.378 10^6m, it turns out that a_r = 0.03 m s^{-2}. By comparing this acceleration with gravity, we see that it is much less than gravity.

12.2.2 Coriolis Force (C)

The second Newton's law is valid in an absolute or inertial coordinate system. We determine (measure) air movement in relation to a system that is related to the Earth. The Earth rotates at an angular velocity in a non-inertial (relative) system. So, the illustration of motion in this relative system must be different than in the absolute.

Scalar quantities such as density ρ(x, y, x, t), pressure, etc. are equal in both systems. However, vector variables, such as vector position **R**, velocity **V** = d**R**/dt, and acceleration **a** = d**V**/dt, differ in these systems. Hence, acceleration must be modified to consider the Earth's rotation. Let's look at the system that rotates with the angular velocity, Fig. 12.6.

Vector **A** that is constant in this system rotates when viewed in the absolute system. In the time interval dt, this vector A (which is normal to the vector A and **Ω**) will change for d**A**. Its absolute value is:

$$\left|d\mathbf{A}\right| = Asin\theta\Omega dt \tag{12.25}$$

because θ is the angle between A and Ω. According to this, vector A changes in the absolute system for:

$$\left|\frac{\mathrm{d}\mathbf{A}}{\mathrm{dt}}\right| = \mathrm{A}\Omega\sin\theta \tag{12.26}$$

and it is normal to A and Ω. In vector form this change is:

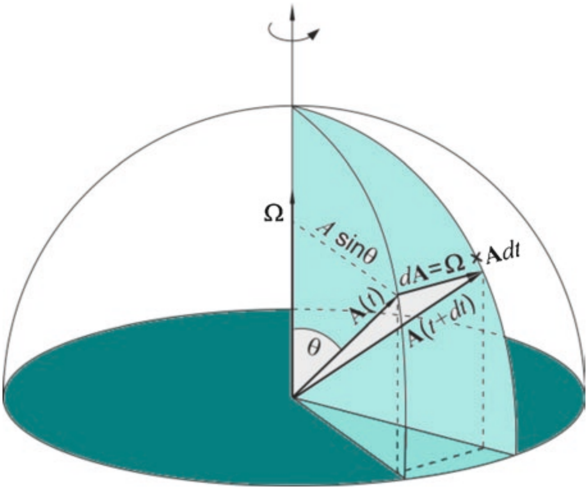


Fig. 12.6 Graphical illustration of Coriolis force

$$\left(\frac{d\mathbf{A}}{dt}\right)_a = \mathbf{\Omega} x \mathbf{A}. \tag{12.27}$$

In general, the individual changes in vector **A** in the absolute system $\left(\frac{d\mathbf{A}}{dt}\right)_a$ and relative system $\left(\frac{d\mathbf{A}}{dt}\right)$ are obviously associated with the following expression:

$$\left(\frac{d\mathbf{A}}{dt}\right)_a = \left(\frac{d\mathbf{A}}{dt}\right) + \mathbf{\Omega} x \mathbf{A} \tag{12.28}$$

If the position vector of the material element of the air is **R**, then its velocity =d**R**/dt , according to (12.28):

$$\mathbf{V}_a = \mathbf{V} + \mathbf{\Omega} x \mathbf{R}. \tag{12.29}$$

Similar applies to acceleration in the absolute system:

$$\left(\frac{d\mathbf{V_a}}{dt}\right)_a = \frac{d\mathbf{V_a}}{dt} + \mathbf{\Omega} x \mathbf{V}_a \tag{12.30}$$

By substituting $\mathbf{V}_a$ from (12.29) in the last expression, we get that:

$$\left(\frac{d\mathbf{V_a}}{dt}\right)_a = \left(\frac{d\mathbf{V}}{dt} + \mathbf{\Omega} x \mathbf{V}\right) + \mathbf{\Omega x}\left(\mathbf{V} + \mathbf{\Omega} x \mathrm{R}\right) = \frac{dV}{dt} + 2\mathbf{\Omega} x \mathbf{V} + \mathbf{\Omega} x\left(\mathbf{\Omega} x \mathbf{R}\right) \tag{12.31}$$

From (12.31) we see that when we want to use relative acceleration instead of the absolute, it will be modified by two additional accelerations. First, 2**Ω** x **V** is Coriolis acceleration. It is conditioned by the individual force that appears in the rotating system and is called the Coriolis force. The Coriolis force **C** that acts on the body of the unit mass is obviously described by the term:

$$\mathbf{C} = -2\mathbf{\Omega} x \mathbf{V} \tag{12.32}$$

Coriolis acceleration is normal to the velocity vector and the planetary vorticity vector 2**Ω**. This acceleration is significant for movements that are by duration (time scale) comparable to the time of rotation of the Earth. The second term in (12.31) is, as we have seen in Sect. 12.3.1, the centrifugal acceleration of air particles due to the Earth's rotation. It should be emphasized once again that when term 2**Ω** x **V** is written together with the acceleration (on the left side of the equation of motion), then it means Coriolis acceleration, and when this term is moved to the right in the equation of motion, it is called the Coriolis force. As this apparent force acts normally on **V**, it does not do any work over the element of air. We see that Coriolis force is completely different from centrifugal force. A centrifugal force occurs when a body rotates, while the Coriolis force occurs only in cases where there is a relative velocity with which the air parcel moves in relation to the rotating system. Therefore, the Coriolis force will always act on air when measured velocities (those obtained by using meteorological instruments) are different from zero.

12.3 Some Common Resultant Forces

Very often there is a need to observe some of the forces or their components together, in terms of their resultant. In general, they represent some complex forces, which under certain conditions significantly affect some scales of air motion. The two most commonly used complex forces are gravity forces and buoyancy forces.

12.3.1 Earth's Gravity Force

From the observed forces that act on motion, we have seen that only the gravity force and the centrifugal force depend on the position of the observed air element relative to the axis of rotation. The resulting force of these two forces is called Earth's gravity or effective gravity, Fig. 12.7. We'll denote it with **G**, so it can be written:

$$\mathbf{G} = \mathbf{G}^* - \Omega\mathrm{x}\left(\Omega\mathrm{x}\mathbf{R}\right) = \mathrm{G}^* + \Omega^2\mathbf{r}. \tag{12.33}$$

This force occurs in a system that rotates with the Earth. Knowing the value of the acceleration that produces forces $\mathbf{G}^*$ and $\mathbf{Z}$ allows us to conclude that $\mathbf{G}^* > Z$. Considering the potentials of the gravitation force Φ' and centrifugal force Φ'', it is obtained:

$$\mathbf{G} = -\nabla\Phi \tag{12.34}$$

where $\Phi = \Phi' + \Phi''$ is the Earth's gravity potential. If in Eq. (12.8) the distance r_p from the centre of the Earth to the point that lies on the MSL in the pole and if in Eq.

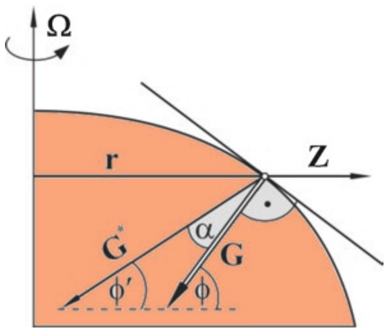


Fig. 12.7 Gravity force **G**

(12.22) is a const = 0, then $\mathbf{\Phi}$ is a *geopotential*. From (12.34) the Earth's gravitational force is expressed in terms of the gradient of the potential $\mathbf{\Phi}$, which is called the geopotential. Since $\mathbf{G} = -g\mathbf{k}$, where $g \equiv |\mathbf{G}|$, it is clear that $\Phi = \Phi(z)$ and $\frac{d\Phi}{dz} = g$. If it is assumed that the geopotential at the MSL must be equal to zero, then the geopotential $\Phi(z)$ at height z is equal to the work that should be exerted against the Earth's gravitation in order to lift the unit mass from MSL to the height, i.e.:

$$\Phi = \int_0^z g dz \tag{12.35}$$

According to this, the geopotential to the surface $\Phi = 0$ coincides with the mean sea level. At the pole $\Phi = 0$.We wonder where else $\Phi = 0$? As the Earth is geoid and that Earth's gravitational force G is always normal at a tangential plane drawn from any point on the Earth's surface, it can be seen by using (12.34) that the geopotential surface $\Phi = 0$ coincides with the Earth's surface. Geopotential surface where $\Phi = 1 = \text{const}$ is found at a greater distance from the Earth's surface on equator than on the pole. This is because G has the highest value at the pole and the smallest on equator. One sees from (12.34) that at the pole since (r = 0) and at the equator (where $G^* \gg Z$) the Earth's gravitation is directed towards the Earth's centre.

12.3.2 Buoyancy Force

We will often be interested in the movement of individual elements of the air in a vertical direction. In this direction, the Earth's gravity and the vertical component of the pressure gradient act as the most significant forces. The vertical component of the pressure gradient force G_{rz} is always pointing upwards (in the positive z-axis direction). The Earth's gravity $\boldsymbol{G}$, i.e. the acceleration it gives to the body of unit mass $\boldsymbol{g}$, is directed in the opposite direction to the z-axis. Let us suppose that the element air of unit mass (air parcel) with density ρ and virtual temperature T_v lies in the environment that is characterized by the following quantities: ρ', T_v', and $\partial p'/\partial z$ (see Fig. 12.8).

Difference

$$S\uparrow = -\frac{1}{\rho}\frac{\partial p'}{\partial z} - g \tag{12.36}$$

is referred as the buoyancy force. Using the static equation, it turns out that:

$$S\uparrow = g\frac{\rho' - \rho}{\rho} \tag{12.37}$$

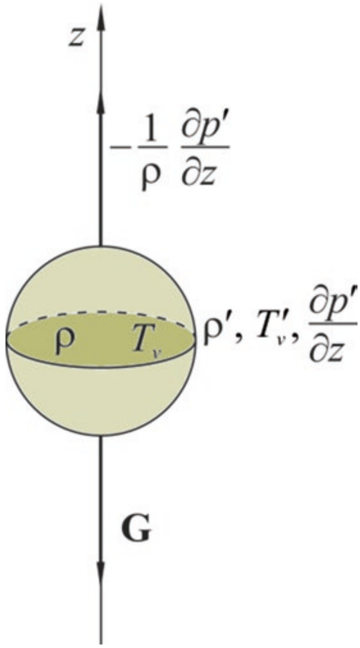


Fig. 12.8 The effect of the buoyancy force and the Earth gravity on the air parcel

We see that the buoyancy force is directed upwards when $\rho' > \rho$ and in the opposite direction when $\rho' < \rho$.

Using the equation of state for moist air, assuming that the pressure in the observed air element is equal to the surrounding pressure at the same height, we obtain that:

$$S \uparrow = g\frac{T_v - T_v'}{\rho T_v'} \tag{12.38}$$

The difference in $T_v - T_v'$ in the atmosphere is usually not large (several degrees). But for certain elements of the air (which are much warmer than the environment due to suitable local conditions), this difference can be large, up to $10^{\circ}C$.
When looking at air unsaturated with water vapour, the expression (12.38) can be written using the potential temperature. It gives:

$$S \uparrow = g\frac{\theta - \theta'}{\theta'} \tag{12.39}$$

The buoyancy force is primarily associated with the stability of the air. When air is heated, it expands and becomes less dense. Since the density changes when changing the temperature, the temperature profile of the layer is very important. If the surface is heated, and if it is colder at altitude, then heated air particles will be able to quickly rise to great altitudes because of the surrounding air temperature which decreases faster than the temperature of warm air particles. This is the basic concept of action under the buoyancy. A practical example of a buoyancy force is shown in Fig. 12.9. In addition, we can calculate the acceleration an unstable air parcel will have and, from this, can determine the parcel's velocity at some later point in time. If $S\uparrow\ > 0$, then the parcel rises; if $S\uparrow\ < 0$, then the parcel descends. We look at the instability at each point in the environmental temperature profile and can determine γ_{env} for each point.

Thus:

$$\mathrm{T}_{v\_env} = \mathrm{T}_0 + \left.\frac{\partial T}{\partial z}\right|_{env} \Delta z = \mathrm{T}_0 - \gamma_{env}\Delta z; \tag{12.40}$$

$$\mathrm{T}_{v\_parcel} = \mathrm{T}_0 + \left.\frac{\partial T}{\partial z}\right|_{parcel} \Delta z = \mathrm{T}_0 - \gamma_{d}\Delta z; \tag{12.41}$$

so that:

$$S\uparrow = \frac{\left(T_0 - \gamma_d \Delta z - T_0 + \gamma_{env}\Delta z\right)}{T_v} = \frac{(\gamma_{env} - \gamma_{d})}{T_v} g\Delta z \tag{12.42}$$

If $\gamma_{env} < \gamma_d$, the parcel accelerates downwards for positive Δz (positive stability);
If $\gamma_{env} > \gamma_d$, the parcel accelerates upwards for positive Δz (negative static stability).

We can also express buoyancy in terms of potential temperature as the nearly conservative variable for adiabatic atmosphere:

Fig. 12.9 The buoyancy effect on a hot air balloon

$$\theta = T\left(\frac{p_0}{p}\right)^{\frac{R_d}{c_p}} \tag{12.43}$$

We want to find $\frac{d\theta}{dz}$. Taking the log of both sides of the equation and replacing a $\frac{dp}{dz} = -g\rho$, we are able to find the following expression for buoyancy in terms of potential temperature:

$$S\uparrow = -g\Delta z\frac{1}{\theta}\frac{d\theta}{dz} \tag{12.44}$$

It does not matter what the environmental temperature or potential temperature profiles; a change in height of an air parcel will result in a temperature that changes along the dry adiabatic and a potential temperature that does not change at all. As you can see below, the stability of a layer depends on the change in environmental potential temperature with height. Air parcels try to move vertically with constant potential temperature. Parcels will move to an altitude (and air density) for which $S\uparrow = 0$. However, if they still have a velocity when they reach that altitude, they will overshoot, experience a negative acceleration, and then descend, overshooting the neutral level again. In this way, the air parcel will oscillate until its oscillation is finally damped out by friction and dissipation of the air parcel. Note that in the neutral section of vertical profile where potential temperature does not change, it is not possible to determine if an air parcel will be stable or unstable. For instance, if the air parcel in the neutral region is given a small upward push, it will continue to rise until it reaches a stable region.

12.4 Atmospheric Motion

In this chapter we will describe the movement of the air under the influence of certain forces. We will also describe some of the terms that are encountered in meteorology and are related to the air movement. Here, we will analyse the effects of the action of individual forces on air movement, which means that we will observe simple atmospheric movements that occur under very limited condition (Zdunkowski and Bott 2003).

12.4.1 The Equation of Motion in a System Rotating Together with the Earth

Since we described the basic forces, we can move on derivation of the equation of motion in a system that rotates together with Earth. The Newton's law of motion can be written in the form:

$$\left(\frac{d\vec{V}_a}{dt}\right) = \sum S \tag{12.45}$$

where a term on the left-hand side of the equation represents an absolute acceleration and on the right-hand side is the sum of the real forces acting on the air parcel of the unit mass.

By substitution of all terms related to the real forces, we obtain:

$$\frac{d\vec{V}}{dt} + 2\vec{\Omega}x\vec{V} + \vec{\Omega}x\left(\vec{\Omega}x\vec{R}\right) - \frac{1}{\rho}\nabla p + \vec{G}^* + \vec{F} \tag{12.46}$$

Here $\vec{V}$ is the relative velocity of air or simply air velocity. Considering that:

$$\vec{G} = \vec{G}^* - \vec{\Omega}x\left(\vec{\Omega}x\vec{R}\right) = \vec{G}^* + \Omega^2\vec{r} \tag{12.47}$$

is the resultant force (Earth's gravitation or effective gravitation) which happens in the system rotating with the Earth, then it follows that Eq. (12.47) is:

$$\frac{d\vec{V}}{dt} = -\frac{1}{\rho}\nabla p + \vec{G} - 2\Omega x\vec{V} + \vec{F} \tag{12.48}$$

In such system, vectors which appear in the vector equation of motion Eq. (12.48) have the following components along x, y, and z coordinates:

$$\frac{d\vec{V}}{dt} = \left(\frac{du}{dt}, \frac{dv}{dt}, \frac{dz}{dt}\right);$$

$$-\frac{1}{\rho}\nabla p = \left(-\frac{1}{\rho}\frac{\partial p}{\partial x}, -\frac{1}{\rho}\frac{\partial p}{\partial y}, -\frac{1}{\rho}\frac{\partial p}{\partial z}\right);$$

$$\vec{G} = \left(0, 0, -g\right); \vec{F} = \left(F_x, F_y, F_z\right)$$

Force **F**, in general, has a component, and in the direction of the z-axis, $\mathbf{F_z}$, although for simplicity, we did not consider in the previous sections. It is evident from Fig. 12.10 that the Ω has components: $\boldsymbol{\Omega} = (\Omega_x, \Omega_y, \Omega_z)$, where $\Omega_x = 0$, $\Omega_y = \Omega \cos\phi$, $\Omega_z = \Omega \sin\phi$ (ϕ) is geographical latitude. Then the Coriolis force obtains the following matrix form:

$$\mathbf{C} = -\mathbf{2}\Omega x\mathbf{v} = -2\begin{vmatrix} \mathbf{i} & j & \mathbf{k} \\ 0 & \Omega_y & \Omega_z \\ u & v & w \end{vmatrix} = \begin{vmatrix} \mathbf{i} & \mathbf{j} & \mathbf{k} \\ 0 & -2\Omega\cos\phi & -2\Omega\sin\phi \\ u & v & w \end{vmatrix} \tag{12.49}$$

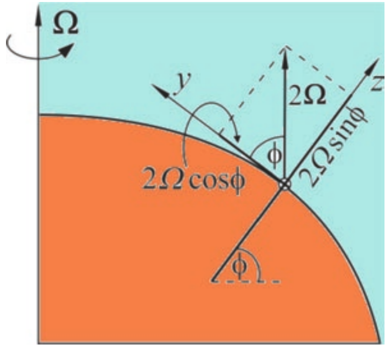


Fig. 12.10 Earth vorticity vector components in (x, y, z) system

$$\text{i.e.}\mathbf{C} = \left(\text{fv} - \text{f}'\text{w}, -\text{fu}, \text{f}'\text{u}\right) \tag{12.49'}$$

with the following parameters being introduced:

$$\text{f} = 2\Omega \sin\phi \text{ and } \text{f}' = 2\Omega \cos\phi \tag{12.50}$$

The parameter f is called Coriolis parameter or the Earth's rotation. Taking in consideration all written before the vector, Eq. (12.48) can be expressed into a system of the following three scalar equations:

$$\begin{aligned}
\frac{du}{dt} &= -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv - f'w + F_x \\
\frac{dv}{dt} &= -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu + F_y \\
\frac{dw}{dt} &= -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + f'u + F_z
\end{aligned} \tag{12.51}$$

In (Eq.12.48) it was not important how the axes (x, y, and z) were oriented. Scalar equations of motion (12.51) are derived under defined directions of axis (x towards the east, y towards the north, and z vertically upwards, normal to the Earth's surface). Therefore, the equation of motion (12.48) can now be written in vector form with defined axes. From (12.51) it follows that:

$$\frac{\text{d}\mathbf{v}}{\text{dt}} = -\frac{1}{\rho}\nabla\text{p} - \text{g}\mathbf{k} - 2\Omega\text{x}\mathbf{v} + \mathbf{F} \tag{12.52}$$

where k is ort in direction of z-axis and g is the acceleration of the Earth's gravity, which is often called effective gravity. The first two equations in (12.51) describe the motion in the plane (x, y) that tangles the Earth at the point of the coordinate origin. Therefore, they are called horizontal equations of motion in the tangential plane. We see that the tangential plane deviates from the surface of the Earth, the more we are moving away from the coordinate origin. This means that horizontal air flows of large proportions (the order of the radius of the radius of the Earth) will not be able to accurately describe this system of equations. In order to avoid this lack, a spherical coordinate system is used. A more detailed information regarding the transformation of this system of equations in spherical coordinate system can be found in several books which deal with dynamic meteorology. However, we will continue to use a system that is often called the system of primitive equations because in further interpretations we will not observe the movement of such large-scale processes as planetary waves. We will mention here that in the system (12.51) we figured out five dependent variables: (u, v, w, p, ρ). Since **g** and Ω are known, F is expressed through (u, v, w). To keep the system of equations closed (means to be completely resolved), additional equations are needed, and we will discuss them later.

12.5 Application of the Equations of Horizontal Motion

12.5.1 Geostrophic Wind ($\mathbf{V}_g$)

In the upper atmosphere above the layer of friction, the friction force which acts on the air particle is practically negligible. Hence, the air particle forces two actions: Coriolis force and the pressure gradient force. When these two forces are in balance with each other, it is called geostrophic balance. Geostrophic wind is therefore uniform horizontal velocity that is balanced with horizontal components of the pressure gradient force and Coriolis force. The graphic illustration of the geostrophic wind flow is given in Fig. 12.11. The direction of movement should be exactly parallel with isobars with high pressure on the right and low pressure on the left. Flow of wind in the upper level is approximately geostrophic. Geostrophic wind (V_g) is a close approximation of the real wind (V). It is important to recall that when an air particle moves, Coriolis force and power of the gradient of pressure change (when the width and pressure change), and this may result in acceleration or deceleration. We will approach the derivation of the equation for the geostrophic wind. From the horizontal momentum equations:

$$\frac{d\mathbf{V}}{dt} = -\frac{1}{\rho}\nabla_h p - f\mathbf{k} \times \mathbf{V}, \tag{12.53}$$

where $\mathbf{V} = (u, v, 0)$, as $\frac{d\mathbf{V}}{dt} = 0$, we have the following term:

$$-\frac{1}{\rho}\nabla_h p - f\mathbf{k} \times \mathbf{V} = 0, \tag{12.54}$$

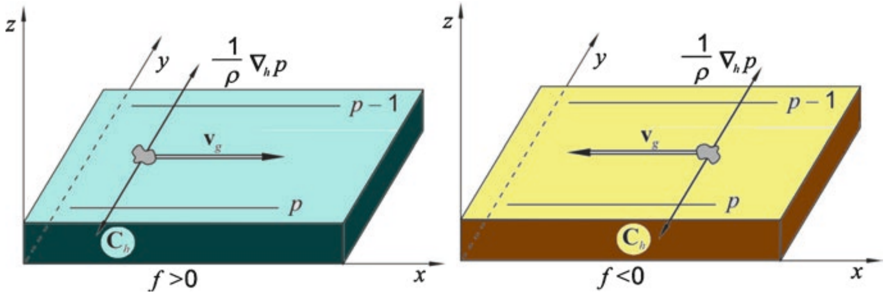


Fig. 12.11 Geostrophic wind at both hemispheres

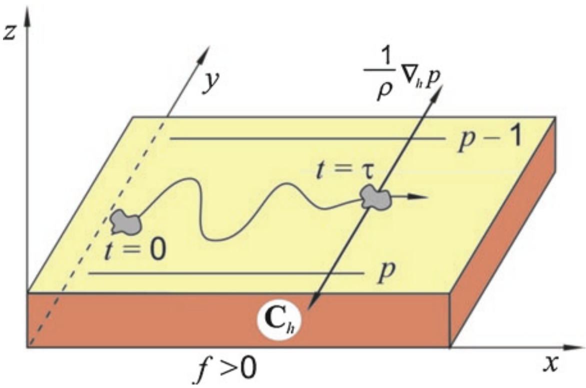


Fig. 12.12 Establishment of the geostrophic equilibrium

Geostrophic wind is a vector ***V*** that satisfies Eq. (12.54). Let us denote it with $\mathbf{V} = \mathbf{V_g}$**.** By vector product of (12.54) with a unit ort "**k**", we obtain that:

$$\mathbf{V_g} = \frac{1}{f\rho}\mathbf{k}\,\mathrm{x}\nabla_{\mathrm{h}}\mathrm{p}, \tag{12.55}$$

which is the vector way of writing the geostrophic wind in the system (x, y, z). The geostrophic wind is therefore a uniform horizontal velocity for which the horizontal component of the pressure gradient force and the Coriolis force is balanced, which is graphically presented in Fig. 12.12. In the atmosphere, in the absence of a frictional force, there is a geostrophic flow. As an illustration for this, we observe the following example: let a gradient of pressure be established at a given moment as it is shown in Fig. 12.13. Then the individual air parcel that was initially in rest will accelerate towards the low-pressure area. Since a particle then has some velocity, normal on the isobar, the Coriolis force $\mathbf{C_h}$ will start to act. When the horizontal pressure gradient force $-\frac{1}{\rho}\nabla_h \mathrm{p}$ and the Coriolis force $\mathbf{C_h}$ are balanced, the movement will be parallel to the isobars. The intensity of geostrophic wind is:

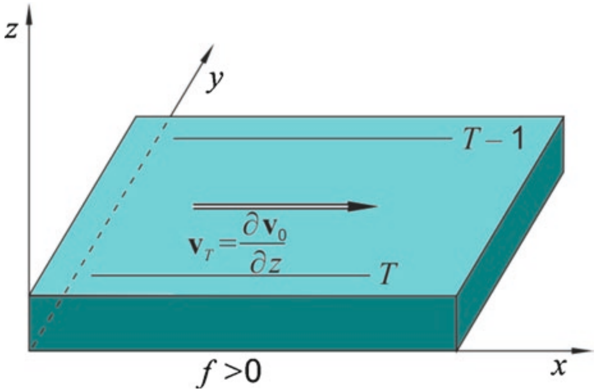


Fig. 12.13 Thermal wind

$$V_g = \frac{1}{f\rho}\frac{\partial p}{\partial n}, \tag{12.56}$$

where n is direction normal to the isobars.

The horizontal momentum Eq. (12.53) can be also written in (p-system) as:

$$\frac{d\mathbf{v}}{dt} = \nabla_p \Phi - f\,\mathbf{k}\,x\,\mathbf{V} \tag{12.57}$$

or absence of friction:

$$0 = \nabla_p \Phi - f\,\mathbf{k}\,x\,\mathbf{V}$$

Hence, we get that:

$$\mathbf{V_g} = \frac{1}{f}\mathbf{k}\,x\,\nabla_p \Phi, \tag{12.58}$$

or the intensity:

$$v_g = \frac{1}{f}\frac{\partial \Phi}{\partial n} \tag{12.59}$$

Here, n is a normal to the isohypse of isobaric surface p=const. The advantage of isobaric coordinate (p-system) is evident from (12.59) as there is no density in this relation. Thus, only the geopotential gradient uniquely determines the geostrophic wind at all heights, while the given pressure gradient corresponds to the different values of the geostrophic wind, depending on the density at different heights.

From (12.58) and (12.59), we see that the geostrophic wind can be calculated, whether we know the pressure field at a certain height or the geopotential field on a fixed isobaric surface. The question is what does this wind represent? The real wind field at altitude is obtained from upper air sounding data. Although the wind is known only in discrete points, we can assume that the vector field is known.

From the assumptions introduced to get the geostrophic wind $\mathbf{V_g}$ from the equation for horizontal motion, we see that $\mathbf{V_g}$ would be equal to real wind in areas where the isobars, i.e. the isohypses, would be parallel and straight and when there would be no frictional force. A scale analysis of the equations of motion shows that assumptions are valid and with great accuracy.

Above the planetary boundary layer, the acceleration and friction force are significantly smaller in magnitude than the pressure gradient force and the Coriolis force. As summarized, the geostrophic wind is a theoretical wind, which is often used to describe the current flow of wind at altitude. Theoretical geostrophic wind is sometimes the same as this wind. However, the only way to obtain values for this wind is the wind to be measured with profilers by using LIDARs and similar instruments or devices. Geostrophic wind is not always uniform throughout the layer. The change of $\mathbf{V_g}$ in height gives the thermal wind $\mathbf{V_T}$.

12.5.2 Thermal Wind (V_T)

By definition, thermal wind $\mathbf{V_T}$ is:

$$\mathbf{V}_\mathrm{T} = \frac{\partial \mathbf{V}_\mathrm{g}}{\partial z} \text{or } \mathbf{V}_\mathrm{T} = \frac{\partial \mathbf{V}_\mathrm{g}}{\partial z} \Delta z. \tag{12.60}$$

Considering (12.60), (12.54), and the equation of the state, we get that:

$$\mathbf{V}_\mathrm{T} = \frac{\partial \mathbf{V}_\mathrm{g}}{\partial z} = \mathbf{V}_\mathrm{T} = \frac{\mathbf{R}}{\mathbf{f}} \frac{\partial T}{\partial z} \mathbf{k} \mathrm{x} \nabla_\mathrm{h} \left(\mathrm{lnp}\right) + \frac{\mathbf{RT}}{\mathbf{f}} \mathbf{k} \mathrm{x} \nabla_\mathrm{h} \left(\frac{\partial}{\partial Z} \mathrm{lnp}\right). \tag{12.61}$$

As the term $\frac{\partial}{\partial Z}\mathrm{lnp} = \frac{1}{\mathrm{p}}\frac{\partial p}{\partial Z} = -\frac{\mathrm{g}}{\mathrm{RT}}$, having in mind (12.54), from the previous term, it is obtained:

$$\mathbf{V}_\mathrm{T} = \frac{\partial \mathbf{V}_\mathrm{g}}{\partial z} = \frac{1}{\mathrm{T}} \frac{\partial T}{\partial Z} \mathbf{V}_\mathrm{g} + \frac{\mathrm{g}}{\mathrm{fT}} \mathbf{k} \mathrm{x} \nabla_\mathrm{h} \mathrm{T} \tag{12.62}$$

The first term on the right side is one order of magnitude smaller than other. One sees that in isotherm atmosphere, the thermal wind is in the same position in relation to isotherms such as the geostrophic wind in relation to the isobars. From the equation for geostrophic wind in *p*-system (12.58), we have:

$$\frac{\partial \mathbf{V}_g}{\partial p} = \frac{g}{f}\mathbf{k} x \nabla_p \frac{\partial z}{\partial p} \tag{12.63}$$

after using the static and the state equation:

$$\frac{\partial \mathbf{V}_g}{\partial p} = -\frac{R}{fp}\mathbf{k} x \nabla_p T. \tag{12.64}$$

By substitution of p from the static equation, the former term gets the form:

$$\mathbf{v_T} = \frac{\partial \mathbf{V}_g}{\partial z} = \frac{g}{fT}\mathbf{k} x \nabla_p T. \tag{12.65}$$

In the p-system, therefore, the thermal wind is completely related to isotherms, as a geostrophic wind relative to the isobars. In Fig. 12.13 are drawn isotherms and the thermal wind on the surfaces p =const.

The introduced term, thermal wind, has a great practical benefit: so, if we know the change of geostrophic wind with height in some layer of atmosphere, from (12.65) we can calculate the horizontal temperature ascendant. Conversely, if we know the geostrophic wind at some height and the horizontal component of the temperature ascendant, we can calculate the geostrophic wind at other altitude. The local temperature change due to horizontal advection can also be calculated using a thermal wind. From:

$$\frac{dT}{dt} = \frac{\partial T}{\partial t} + \mathbf{v}_g \cdot \nabla_p T \tag{12.66}$$

Since we consider the case when $\frac{dT}{dt} = 0$, as the individual air parcels do not change the temperature but only move in the horizontal direction at velocity $\mathbf{V}_g$, we get:

$$\frac{\partial T}{\partial t} = -\mathbf{V}_g \cdot \nabla_p T \tag{12.67}$$

By vector product of (12.65) with unit ort vector in z-direction **k**, we have:

$$\nabla_p T = -\frac{f}{g} T\mathbf{k} x \frac{\partial \mathbf{V}_g}{\partial z} \tag{12.68}$$

this in turn into (12.67) gives:

$$\frac{\partial T}{\partial t}=\frac{f}{g}T\mathbf{V}_g\cdot\left(k\,x\frac{\partial \mathbf{V}_g}{\partial z}\right). \tag{12.69}$$

In accordance with the method of writing a mixed product, the previous term is more convenient to write in the form:

$$\frac{\partial T}{\partial t}=-\frac{f}{g}T\mathbf{k}\cdot\left(\mathbf{V}_g\,x\frac{\partial \mathbf{V}_g}{\partial z}\right). \tag{12.70}$$

We see if $\left(\mathbf{V}_g\,x\frac{\partial \mathbf{V}_g}{\partial z}\right)$ is directed opposite from the unit vector **k**, so $\frac{\partial T}{\partial t}>0$ at $f>0$, i.e. warm advection will be performed. Such a situation is shown in Fig. 12.14 (left panel).

If $\left(\mathbf{V}_g\,x\frac{\partial \mathbf{V}_g}{\partial z}\right)$ is in the direction of the vector **k,** as it is shown in Fig. 12.14 (right panel), then so $\frac{\partial T}{\partial t}<0$, and we say that cold advection is then performed. The temperature change due to advection does not exist when $\mathbf{V}_g$ and $\frac{\partial \mathbf{V}_g}{\partial z}$ are of the same direction. When $\alpha = 90$, the temperature change is the highest. In practical procedures for calculating the local temperature change from (12.70), we need to represent $\frac{\partial \mathbf{V}_g}{\partial z}$ in terms of the final differences (since we do not know the function $\mathbf{V}_g = \mathbf{V}_g(z)$). In that case, it is:

$$\frac{\partial \mathbf{V}_g}{\partial z}\cong\frac{1}{z_2-z_1}\left(\mathbf{V}_{g2}-\mathbf{V}_{g1}\right) \tag{12.71}$$

where $z_2 - z_1$ is the atmospheric layer depth between the levels where $\mathbf{V}_g = \mathbf{V}_{g2}$ and $\mathbf{V}_g = \mathbf{V}_{g1}$. Then (12.70) get the following form (using $\mathbf{V}_g = \mathbf{V}_{g1}$):

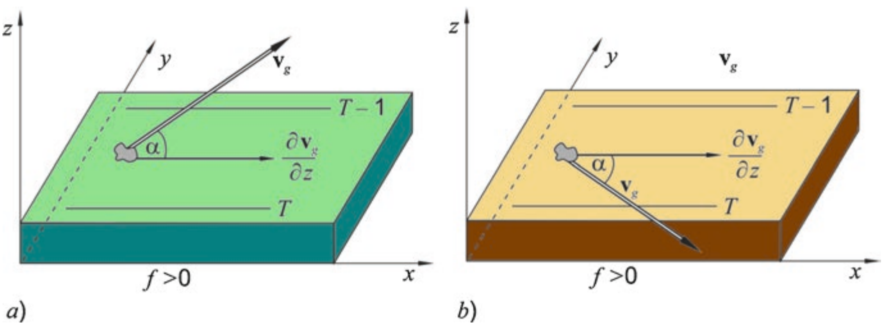


Fig. 12.14 Warm and cold advection

$$\frac{\partial T}{\partial t} = -\frac{\mathrm{fT}}{\mathrm{g}\left(\mathrm{z}_2 - \mathrm{z}_1\right)} \mathbf{k} \cdot \left(\mathbf{V}_{\mathrm{g}1} \mathrm{x} \mathbf{V}_{\mathrm{g}2}\right) \tag{12.72}$$

In the relation (12.72), term $\mathbf{k} \bullet (\mathbf{V}_{g1} x\ \mathbf{V}_{g2})$ is quantitatively equal to the surface of parallelogram which sides $\mathbf{V}_{g1}$and $\mathbf{V}_{g2}$. Thus, it follows if $z_2 - z_1 = 1$ (for unique change of height):

$$\frac{\partial T}{\partial t} = -\frac{2\mathrm{f}}{\mathrm{g}} \mathrm{T} \frac{\mathrm{dS}}{\mathrm{dz}}. \tag{12.73}$$

where

$$\frac{\mathrm{dS}}{\mathrm{dz}} = \frac{1}{2} \mathbf{k} \cdot \left(\mathbf{V}_{\mathrm{g}} \mathrm{x} \frac{\partial \mathbf{V}_{\mathrm{g}}}{\partial z} \right). \tag{12.74}$$

It is assumed that $S > 0$ when $\mathbf{V}_g$ moves to the left from $\mathbf{V}_{g1}$ to $\mathbf{V}_{g2}$.

In barotropic atmosphere, as $\rho = \rho(p)$, i.e. at the surfaces $p=const.$, it follows that and density is constant. From the equation of state for dry air, it follows that $\nabla_h T = 0$. Hence, it follows that the geostrophic wind does not change with the height in the barotropic atmosphere. From there it follows that the local temperature change, due to horizontal advection, is then zero.

12.5.3 Gradient Wind

Looking at the movement of air on weather maps, a great complexity can be noticed. Nevertheless, the field of pressure (or geopotential) and wind field can be associated with the assumption that there is a balance of forces acting on the individual air parcel. In order to better understand the balance of forces, the movement is idealized. It is assumed that the motion is stationary (not dependent on time), horizontal, and carried out in the absence of frictional force. Such simplified movement conditions, such as geostrophic motion, can only be achieved in special circumstances. Isobars (isohypses) are rarely straightforward. Horizontal frictionless motion, parallel to such isobars (isohypses), when it is non-accelerated along the isobars (tangential acceleration $\frac{d\mathbf{V}}{dt} = 0$, but the radial acceleration $V^2/r \neq 0$) is more realistic than geostrophic. It is called *gradient motion*. In such motion there is a balance between the three forces: the horizontal components of the pressure gradient force p, the Coriolis force C_h, and the centrifugal force C_c. From the equations with a horizontal motion written in a natural coordinate system, and the air parcel path along the curved isobar is denoted with $\boldsymbol{s}$, remains the expression:

$$\frac{v^2}{r} + fv + \frac{1}{\rho}\frac{\partial p}{\partial n} = 0 \tag{12.75}$$

Velocity calculated from (12.75) is known as gradient velocity (or gradient wind). Let's denote it with V_{gr}. It is obvious:

$$V_{gr} = -\frac{fr}{2} \pm \sqrt{\frac{f^2 r^2}{4} - \frac{r}{\rho}\frac{\partial p}{\partial n}} \tag{12.76}$$

It is interesting to observe a special case of a circular motion of air in a cyclone, when the pressure gradient force is balanced with a centrifugal force. It is such a movement where the Coriolis force is considerably smaller than the others individually. Then Rossby's number $R_O = \frac{v^2}{r} / fV$ is very large. From the balance of forces, it follows that:

$$\frac{V^2}{r} = -\frac{1}{\rho}\frac{\partial p}{\partial n} \tag{12.77}$$

Hence:

$$V = \sqrt{-\frac{r}{\rho}\frac{\partial p}{\partial n}} \tag{12.78}$$

With expression (12.78) so-called cyclostrophic velocity is defined. It can be directed in both positive and negative senses. Horizontal air motion near the vertical axis of rotation, in the case of hurricanes, is very similar to cyclostrophic motion. Typical mid-latitude motions, unlike the previous one, are like geostrophic motion, i.e. there is a balance between pressure gradient force and Coriolis force. The relationship between the defined geostrophic and gradient wind in certain, identical conditions can be easily found. Geostrophic velocity in a natural coordinate system:

$$V_g = -\frac{1}{f\rho}\frac{\partial p}{\partial n}$$

Taking this into account, (12.75) can be written in the form:

$$fV_{gr} + \frac{V_{gr}^2}{r} - fV_g = 0. \tag{12.79}$$

If we multiply (12.75) by $\left(-\frac{1}{V_{gr}^2}\right)$, we can also write as:

$fV_g\left(\frac{1}{V_{gr}}\right)^2 - f\frac{1}{V_{gr}} - \frac{1}{r} = 0$, where the solution is:

$\frac{1}{V_{gr}} = \frac{f \pm \sqrt{f^2 + \frac{4fV_g}{r}}}{2fV_g}$. Hence, we get:

$$\frac{V_g}{V_{gr}} = \frac{1}{2} \pm \sqrt{\frac{1}{4} + \frac{V_g}{rf}} \tag{12.80}$$

From (12.80) we see that in front of the root, only positive sign should be used, since as $r \to \infty$ follows that $V_g = V_{gr}$, which is possible only in case of a positive sign.

When $f > 0$ and $r > 0$ (cyclonic circulation), it follows that $V_g > V_{gr}$.

When $f > 0$ and $r < 0$ (anticyclonic circulation), it follows that $V_g < V_{gr}$.

From the previous condition, one sees that the geostrophic wind in the cyclonic curves of the isobars is greater than the speed at which there is a balance of forces for this form of the isobars. In anti-cyclonic case, the geostrophic wind has a lower value than the velocity at which a balance of forces exists for this form of isobars.

12.5.4 Quasi-geostrophic Equations of Motion

The scalar equation of horizontal momentum equations into $p - system$ written in the vector form is:

$$\frac{d\mathbf{V}}{dt} = -\nabla_p \Phi - f\,\mathbf{k} \mathrm{x} \mathbf{v} \tag{12.81}$$

where $\mathbf{V} = u\mathbf{i} + v\mathbf{j}$ is velocity vector (horizontal component of the air velocity vector) and ∇_p is horizontal operator that is applied on isobaric surface. As we have seen, this represents a simplified equation, due to the application of the hydrostatic approximation. That still contains some terms which are less significant (secondary importance) for synoptic-scale motions at mid-latitudes. Therefore, it will be simplified by introducing the fact that the horizontal motion is approximately geostrophic.

If we denote the wind vector as:

$$\mathbf{V} = \mathbf{V_g} + \mathbf{V_a} \tag{12.82}$$

where the geostrophic wind $\mathbf{V_g}$ is defined by the expression:

$$\mathbf{V_g} = \frac{1}{f_0} \mathbf{k} \mathrm{x} \nabla \Phi. \tag{12.83}$$

In (12.82) $\mathbf{V_a}$ is a difference between the real wind vector and the geostrophic wind vector and is referred as *geostrophic wind***.** In (12.83) an assumption is introduced that the meridional scale of motions L is small relative to the Earth's radius a, so the geostrophic wind is expressed with aid of the constant value of the Coriolis parameter f_0. For the observed motions $|\mathbf{V_g}| \gg |\mathbf{V_a}|$**,** it is apparent that the interpretation of ageostrophic wind depends on that whether the geostrophic wind is defined by f_0 (12.83) or a variable Coriolis parameter f (12.58). In the first case, geostrophic wind is non-divergent, while in the second case, geostrophic wind has divergent part. Although the geostrophic wind can be defined with f_0, it is necessary in this expression to maintain the dependence of the Coriolis parameter on the latitude. Keeping only the first two terms follows that:

$$f = f_0 + \beta y \tag{12.84}$$

where $\beta = \frac{df}{dy}\phi_0 = 2\Omega \cos\phi_0 / a$ and it is taken that $y = 0$ at $\phi_1 = \phi_0$. This approximation is called *β-plane approximation*. For the synoptic scale of motions, the relationship between the terms of the r.h.s. of (11.1.3) is:

$$\frac{\beta L}{f_0} \sim \frac{\cos\phi_0 L}{\sin\phi_0 a} \ll 1.$$

By this using f_0 instead of the real value of the Coriolis parameter f is approved.

From (12.81) it is evident that the acceleration, following the motion, is equal to the difference between the Coriolis force and the pressure gradient force. This difference depends on the deviation of the real wind from geostrophic wind. So, it is not really appropriate to simply replace the horizontal velocity with the geostrophic, so it is not justified simply to replace the horizontal velocity with a geostrophic value in a Coriolis term, but by replacing (12.83), (12.82), and (12.84) in the terms on the r.h.s. of (12.81), it is obtained:

$$\mathrm{f}\,\mathbf{k}\,\mathrm{x}\,\mathbf{V} + \nabla\Phi = (f_0 + \beta y)\mathbf{k}\,\mathrm{x}\left(\mathbf{V_g} + \mathbf{V_a}\right) - f_0\mathbf{k}\mathrm{x}\mathbf{V}_g \approx f_0\mathbf{k}\mathrm{x}\mathbf{v}_a + \beta y\mathbf{k}\,\mathrm{x}\,\mathbf{V}_g \tag{12.85}$$

Here, in terms proportional with βy, the ageostrophic wind is ignored relative to the geostrophic wind. Thus, the approximate equation of motion is:

$$\frac{\mathrm{d}\mathbf{V}_g}{\mathrm{dt}} = -f_0\,\mathbf{k}\,\mathrm{x}\,\mathbf{V_a} - \beta y\mathbf{k}\,\mathrm{x}\,\mathbf{V}_g \tag{12.86}$$

It is called a **quasi-geostrophic equation** of motion. It should be noted that here the operator for an individual (substantial) derivative is defined as:

$$\frac{\mathrm{d}}{\mathrm{dt}} = \frac{\partial}{\partial t} + \mathbf{V}_g \cdot \nabla = \frac{\partial}{\partial t} + u_g\frac{\partial}{\partial x} + v_g\frac{\partial}{\partial y} \tag{12.87}$$

12.6 Vertical Motions in the Atmosphere

For the synoptic scale of motion, the vertical velocity is of order a few centimetres per second. By the standard upper air sounding, the wind speed is measured with accuracy of about one metre per second. Therefore, there is no measurement of the vertical velocity into the standard meteorological practice. Usually, the vertical velocity is kinematically calculated (using the continuity equation) and with adiabatic approach (utilizing thermodynamics equation). Both methods are applied in the isobaric (p)-system, where $\omega(p)$ is used instead of $w(z)$. These two quantities of vertical motion will be correlated with aid of the hydrostatic approximation. By definition $\omega = dp/dt$ (so-called omega equation) when the individual derivative is developed into (x, y, z) system, it is obtained:

$$\omega = \frac{dp}{dt} = \frac{\partial p}{\partial t} + \boldsymbol{V} \cdot \nabla p + w\frac{\partial p}{\partial z} \tag{12.88}$$

As we have seen before, the horizontal velocity can be expressed through the geostrophic $\boldsymbol{V}_g$ and ageostrophic velocity, $\boldsymbol{V} = \boldsymbol{V}_g + \boldsymbol{V}_a$, where $|\boldsymbol{V}_g| \gg |\boldsymbol{V}_a|$. In addition, $\boldsymbol{V}_g = \frac{1}{\rho f} kx\nabla p$. Hence, it follows that $\boldsymbol{v}_g \cdot \nabla p = 0$. Considering that and using the hydrostatic approximation, we can write that:

$$\omega = \frac{\partial p}{\partial t} + \boldsymbol{V}_a \cdot \nabla p - g\rho w. \tag{12.89}$$

Three terms at the r.h.s. of (12.89) have the following typical values:

$$\frac{\partial p}{\partial t} \sim \frac{10mb}{per\,day};$$

$$\boldsymbol{V}_a \cdot \nabla p \sim \left(\frac{1m}{s}\right)\left(\frac{0.01mb}{1km}\right) \sim \frac{1mb}{per\,day}; \quad g\rho w \sim \frac{100mb}{day}$$

Therefore, it is seen that is quite appropriate to use this relation in the practice, in order to calculate vertical velocity from the omega equation:

$$\omega = g\rho w. \tag{12.90}$$

12.7 Vorticity Equation

From equations of motion, together with the equation of thermodynamics and continuity equation, the corresponding combinations yield *derived equations*. They contain unknown important quantities such as vorticity, divergence, etc. When in these equations we have time derivatives, they are called prognostic equations. When they do not contain derivatives, they are known as diagnostic equations. In this context, using the following procedure, a vorticity equation is derived. From the momentum equations in the horizontal plane:

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}+w\frac{\partial u}{\partial z}=-\alpha\frac{\partial p}{\partial x}+fv+F_x \tag{12.91}$$

$$\frac{\partial v}{\partial t}+u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}+w\frac{\partial v}{\partial z}=-\alpha\frac{\partial p}{\partial y}-fu+F_y \tag{12.92}$$

by differentiating the first by y and the other by z and subtracting the first from the second of the two differentiated equations, we obtain that:

$$\begin{aligned}&\frac{\partial \zeta}{\partial t}+u\frac{\partial \zeta}{\partial x}+v\frac{\partial \zeta}{\partial y}+w\frac{\partial \zeta}{\partial z}-\left(f+\zeta\right)D\\&-v\frac{\partial f}{\partial y}-\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z}-\frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right)-\left(\frac{\partial \alpha}{\partial x}\frac{\partial p}{\partial y}-\frac{\partial \alpha}{\partial y}\frac{\partial p}{\partial x}\right)+\left(\frac{\partial F_y}{\partial x}-\frac{\partial F_x}{\partial y}\right)\end{aligned} \tag{12.93}$$

where:

$$\zeta=\frac{\partial v}{\partial x}-\frac{\partial u}{\partial y},\quad D=\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}.$$

As f only depends on y, thus it follows:

$$v\frac{\partial f}{\partial y}=\frac{df}{dt}$$

so, the Eq. (12.93) could be written into a more concise form:

$$\begin{aligned}\frac{d}{dt}\left(f+\zeta\right)=-\left(f+\zeta\right)D-\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z}-\frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right)\\-\left(\frac{\partial \alpha}{\partial x}\frac{\partial p}{\partial y}-\frac{\partial \alpha}{\partial y}\frac{\partial p}{\partial x}\right)+\left(\frac{\partial F_y}{\partial x}-\frac{\partial F_x}{\partial y}\right)\end{aligned} \tag{12.94}$$

Equation (12.94) is the vorticity equation. It shows the individual change of the absolute vorticity $\zeta_a=\zeta+f$ (it is reffered to us as the absolute vorticity since f is the vertical component of vorticity of Earth related to fixed system and ζ is vertical component of vorticity in relation to the Earth) which depends on divergence D:

$$D = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \tag{12.95}$$

twisting term P:

$$P = -\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) \tag{12.96}$$

the solenoidal term S is denoted as:

$$S = -\left(\frac{\partial \alpha}{\partial x}\frac{\partial p}{\partial z} - \frac{\partial \alpha}{\partial y}\frac{\partial p}{\partial x}\right) \tag{12.97}$$

and the last term (friction term) which depends on friction F_r:

$$F_r = \left(\frac{\partial F_y}{\partial x} - \frac{\partial F_x}{\partial y}\right) \tag{12.98}$$

It can be easily shown that a twisting term can be expressed as follows:

$$\mathrm{P} = -\left(\frac{\partial w}{\partial x}\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y}\frac{\partial u}{\partial z}\right) = \left(\eta\frac{\partial w}{\partial x} + \xi\frac{\partial w}{\partial y}\right)$$

where:

$$\eta = \frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}; \xi = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}$$

The quantities η and ξ are the horizontal components of a three-dimensional vorticity vector ξ = (η, ξ, ζ).

It is also easily seen that the solenoid term S (Eq. 12.97) can be written in the form:

$$S = -\left(\nabla\alpha x \nabla p\right)\cdot \boldsymbol{k}.$$

Substituting the last two expressions for P and S into (Eq. 12.94) yields that:

$$\frac{d}{dt}\left(f + \zeta\right) = -\left(f + \zeta\right)D - \left(\eta\frac{\partial w}{\partial x} + \xi\frac{\partial w}{\partial y}\right) - = -\left(\nabla\alpha x \nabla p\right)\cdot \boldsymbol{k} \tag{12.99}$$

It can be seen from (Eq. 12.99) that the absolute vorticity of the individual air element changes due to two conditions:

1. If the divergence term is positive ($D > 0$), then the air element stretches over a large area, thus causing the vorticity to decrease in order to conserve the angular momentum.
2. When the divergence term is negative ($D < 0$), then the air element shrinks over small area and elongates. In this case the relative vorticity gradually increases (due to the reduced inertial moment). This is shown in Fig. 12.15. As it is shown in Fig. 12.15a when there is stretching, the divergence term:

$$\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)0=-\left(f+\zeta\right)D=-\left(f+\zeta\right)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)>0$$

$$=>\frac{d}{dt}\left(f+\zeta\right)>0\ \ =>f+\zeta$$

must increase.

In case of compression (Fig. 12.15b), it follows that:

$$\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)>0=>-\left(f+\zeta\right)D=-\left(f+\zeta\right)\left(\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)<0$$

$$=>\frac{d}{dt}\left(f+\zeta\right)0\ \ =f+\zeta$$

must decrease.

1. The twisting term is called because the horizontal shear of the vertical velocity ($\partial w/\partial x$, $\partial w/\partial y$) overturns the element of air and its horizontal component of vorticity (η, ξ,) transforms into a vertical component of vorticity. This can be seen in Fig. 12.16. The strong, vertical shear of horizontal motion, which occurs at high-temperature gradients, enhances the components η and ξ which, after spinning up, represents a strong source for the vertical component of vorticity ζ. Tilting or twisting of horizontal vorticity into the vertical (or vertical vorticity

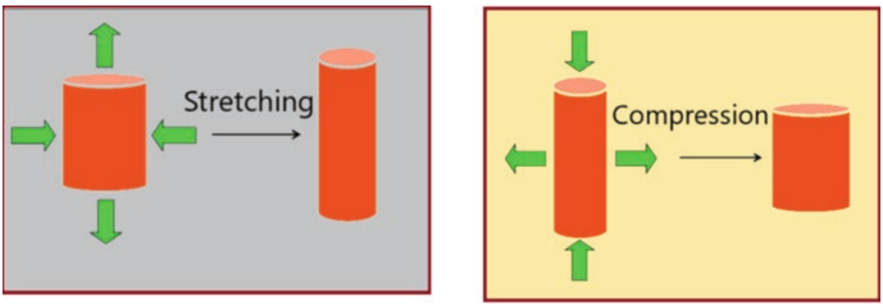


Fig. 12.15 Divergence term. Stretching (left panel) Compression (right panel)

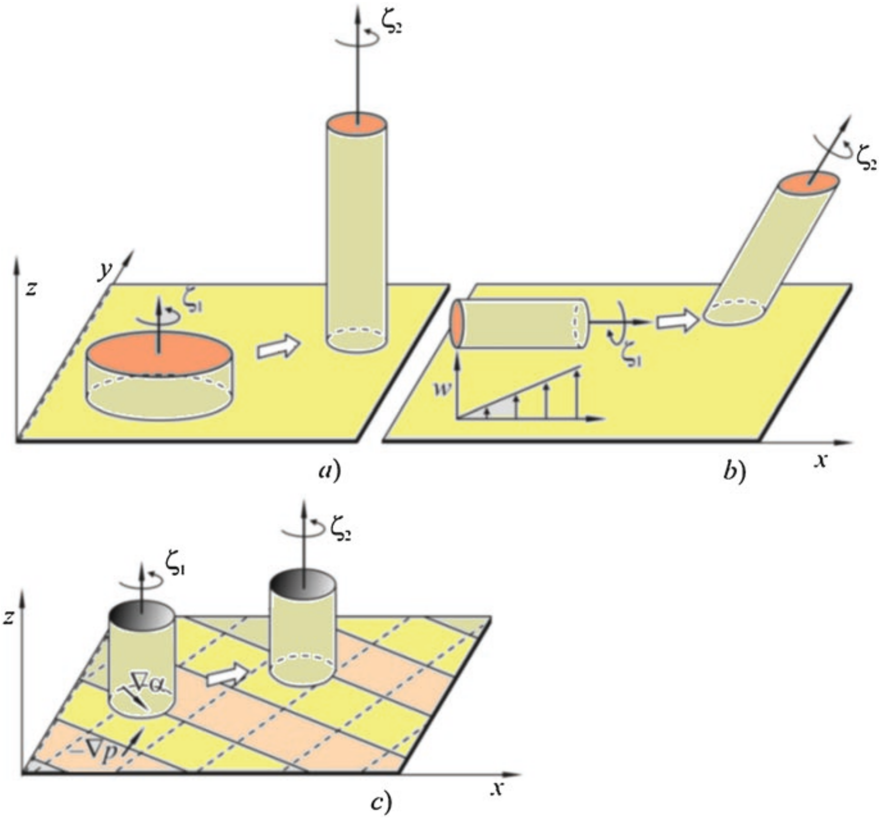


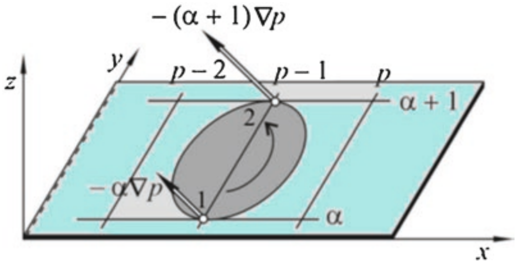
Fig. 12.16 Production of absolute vorticity of the air element

into the horizontal) is important for mesoscale storm dynamics and tornadoes, but not typically a major player in the large scale.

2. The solenoid term, which appears in the baroclinic atmosphere, generates a vertical component of vorticity. In such case pressure and density (i.e. temperature) contours cross. Vorticity changes result from differential horizontal accelerations. As seen in Fig. 12.16b, the pressure gradient force of the observed air element at point (2) is greater than at point (1).
3. Pressure gradient forces create a spinning moment (torque) that changes the angular momentum of an air element occupied by a single solenoid.

Although $\dfrac{\partial p}{\partial y}$ is constant in the illustration above (see Fig. 12.16b), the pressure gradient acceleration is larger on the right because the density is lower. As a result, the relative vorticity gradually increases (i.e. becomes cyclonic if starting from rest). Mathematically the above becomes:

Fig. 12.17 Contribution of isobar-isosteric solenoids of air vorticity



$$\frac{\mathrm{D}}{\mathrm{Dt}}(\mathrm{f}+\zeta)=\frac{1}{\rho^2}\left(\frac{\partial\rho}{\partial x}\frac{\partial p}{\partial y}-\frac{\partial\rho}{\partial y}\frac{\partial p}{\partial x}\right)=\frac{1}{\rho^2}\left(\frac{\partial\rho}{\partial x}\frac{\partial p}{\partial y}\right)>0$$

$$\text{Since}\quad \rho>0,\ \frac{\partial\rho}{\partial x}<0\,\text{and}\ \frac{\partial p}{\partial y}<0$$

The last term on the right side of Eq. (12.94) is due to the existence of a change in the friction force in the horizontal plane. If there is a torque (spinning momentum) caused by friction, the particle acquires cyclonal or anticyclone circulation (see Fig. 12.17). As the result of this:

$$\frac{d}{dt}(f+\zeta)\gtreqless 0. \tag{12.100}$$

12.8 Basic Characteristics of Vorticity

Vorticity by definition is a vector field which defines a microscopic measure of air parcel spin about its axis. As it is discussed in the previous section, there are three types of vorticity: Earth's vorticity, relative vorticity, and absolute vorticity.

Earth vorticity occurs as the result of Earth's rotation. It is also called *planetary vorticity* that is a twice the angular velocity 2Ω. The local vertical component of the planetary vorticity $f = 2\Omega sin\phi$ where Ω is the angular speed of the earth and ϕ is the latitude. The relative vorticity is the vorticity relative to the Earth induced by the air velocity field. This vorticity is induced due to troughs and ridges and horizontal wind speed differences. Since air flow in troughs is cyclonic and in ridges anticyclonic, relative vorticity in troughs is positive and in ridges anticyclonic.

The *absolute vorticity* of the air parcel is the sum of the relative vorticity and Earth vorticity.

Vorticity caused by a change in wind direction or wind speed with height is termed *horizontal vorticity* (the spin is in relation to a horizontal axis) and directional wind shear (veering). Horizontal vorticity is most important in the atmospheric boundary layer (ABL) (lower part of troposphere).

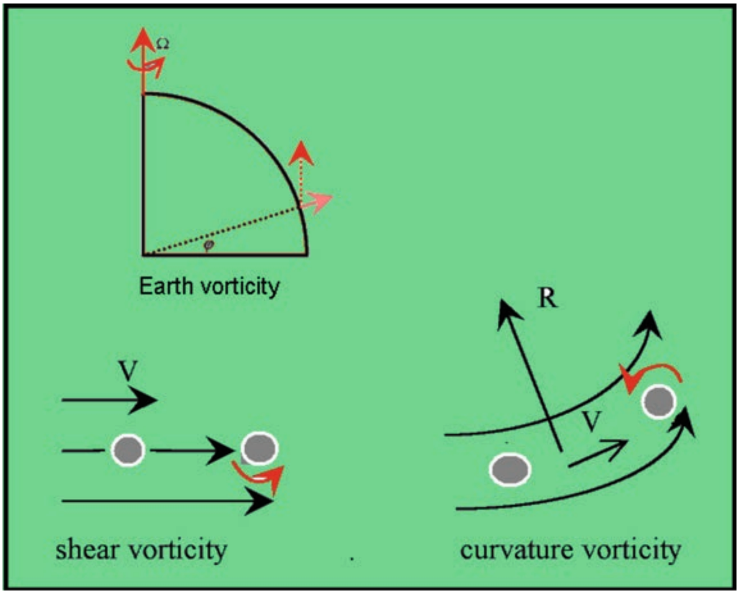


Fig. 12.18 Diagram of a three vorticity components

Streamwise vorticity is the amount of horizontal vorticity that is parallel to storm inflow. Storm inflow is the velocity of the low-level wind moving towards a thunderstorm. Helicity is the amount of streamwise vorticity that is available to be ingested by a thunderstorm. In summary, *vertical positive vorticity* contributes to upper level divergence in the PVA region and thus rising air, while *horizontal vorticity* is important to severe weather. In the synoptic practice, vorticity is evaluated and plotted on the 500 mb chart. Vorticity is a clockwise or counterclockwise spin in the troposphere. A 500 mb vorticity is also termed vertical vorticity. This vorticity is caused by troughs and ridges and other embedded shortwaves or highs. A wind flow through a vorticity gradient will produce regions of PVA (positive vorticity advection) and NVA (negative vorticity advection). PVA contributes to rising air. While low pressure is associated with rising air and updraft motion, high pressure causes sinking air.

In general the vorticity depends on Earth vorticity (the spinning motion created by the Earth's rotation), shear vorticity (wind speed shear), and curvature vorticity (wind veering or directional shear) which elements are shown in Fig. 12.18.

It is obvious that counterclockwise rotation in the Northern Hemisphere results in a positive vorticity, while clockwise rotation produces negative vorticity.

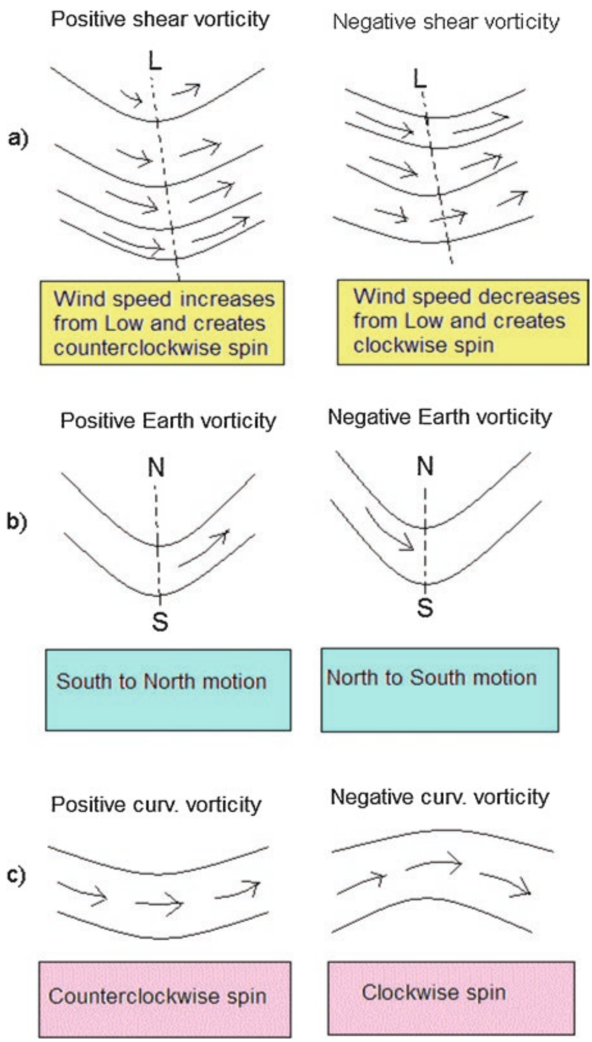


Fig. 12.19 Diagram of a positive and negative vorticity

12.8.1 Positive Vorticity Advection PVA and Upward Air Motion

Positive-increasing vorticity appears **i**n case of a positive shear vorticity the wind velocity increasing when moving away from centre point of trough (left panel in Fig 12.19a). In this case a south to north movement of air exists which means that Coriolis increases when moving from the equator towards the poles (Fig. 12.19b). In addition this occurs in troughs and shortwaves in presence of positive curvature vorticity (left panel in Fig. 12.19c).

Negative-decreasing vorticity is observed in negative shear vorticity under North-South air parcel movement and negative curvature vorticity-clockwise spinning (right pannels in Fig. 12.19).

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Chapter 13
Atmospheric Waves

The atmosphere has a wave nature with various types of wave motions (Lin 2007; Hoskins and James 2014). It represents an environment which allows the formation and maintenance of different waves. Let's first look briefly at the wave physical concept before continuing with the description and derivation of atmospheric waves.

13.1 Waves: General Features

In physics, a wave is simply a disturbance that transfers energy through matter or space. A schematic illustration of a wave is shown in Fig. 13.1, where λ is wavelength, k-vector ($k = 2\pi/\lambda$), κ – wave number $\kappa = (1/\lambda)$, τ is period, ω is angular frequency ($\omega = 2\pi/\tau$) and ν is frequency$(1/\tau)$.

There are two main types of waves: mechanical and electromagnetic (Fig. 13.2). A mechanical wave represents an oscillation of matter transferring energy through a medium. Electromagnetic waves are oscillations of electric and magnetic fields that propagate at the speed of light through a vacuum.

Using the physical concept, we can express the simple wave propagation equation. If we displace any function $f(x)$ to the right and change its argument from x to $x - a$, where a is a positive number, and if we put $a = vt$, where v is positive and t is time, then the displacement will increase with time. Thus $f(x + vt)$ represents a rightward, or forward, propagating wave. Similarly, $f(x - vt)$ represents a leftward, or backward, propagating wave, where v is the wave speed.

According to the usual classification of mechanical waves in the literature, there are the following types of waves: transverse, longitudinal and surface.

A characteristic of transfer waves is that the medium oscillates at a right angle to the direction of wave motion. These waves have two parts: the crest and the trough. Longitudinal waves are waves in which the displacement of the medium is in the same or opposite direction to the direction of propagation of the wave.

V. Spiridonov, M. Curic, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_13

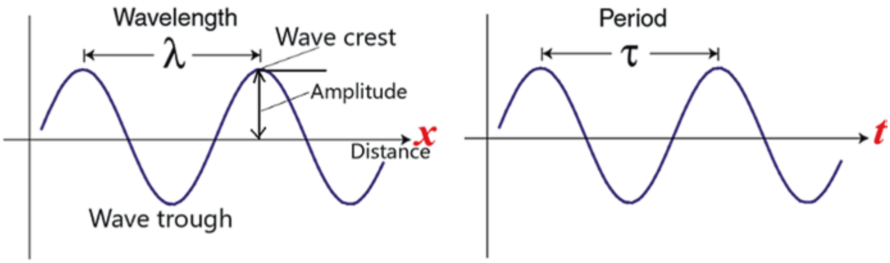

Fig. 13.1 Schematic illustration of wave

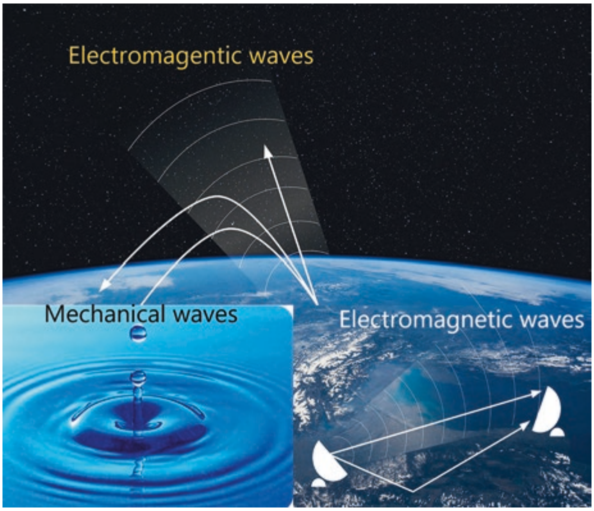

Fig. 13.2 An example of mechanical and electromagnetic waves

A mechanical wave propagating along the boundary region of two different media is called *a surface wave*. Electromagnetic waves do not require a medium. On the contrary, they represent periodic vibrations of the electromagnetic field generated by charged particles, which propagate through the vacuum. These types have a wide range of wavelengths, starting from radio waves (large wavelengths) through microwaves, infrared, visible, ultraviolet radiation to x-rays and gamma rays (shortest waves).

13.2 Wave Equation

The wave equation is an important second-order linear partial differential equation for description of waves such as mechanical (water waves, sound waves and seismic waves) or light waves (acoustics, electromagnetics and fluid dynamics). The 1-D wave equation was first discovered by the French scientist Jean-Baptiste le Rond d'Alembert in 1746, and 10 years later, Euler established the 3-D partial differential wave equation. It typically concerns a time variable t, one or more spatial variables $x_1, x_2, \ldots, x_n$ and the scalar function $u = u(x_1, x_2, \ldots, x_n; t)$ as a useful approximation for modelling the mechanical displacement of waves. It follows that the wave equation for u is

$$\frac{\partial^2 u}{\partial t^2} = c^2 \nabla^2 u \,\text{or}\, \frac{\partial^2 u}{\partial t^2} = c^2 \frac{\partial^2 u}{\partial x^2} \left(1-\text{d wave equation}\right), \tag{13.1}$$

where u is the amplitude of the wave at position x and time t and c is the velocity of the wave. Solutions of this equation describe the propagation of the disturbance from the source region with a constant velocity (c) in one or more directions. The 1-D wave equation for any scalar function (f) can be written as

$$\frac{\partial^2 f}{\partial t^2} - \frac{1}{v^2}\frac{\partial^2 f}{\partial t^2} = 0, \tag{13.2}$$

where (v) is wave speed. The wave equation has the simple solution $f(x, t) = f(x \pm vt)$.

13.2.1 Mathematical Description of 1-D Waves

When a simple harmonic motion propagates in a given direction (for example, the positive x-axis), harmonic waves are created. On that occasion, each point in the medium where the waves propagate achieves a uniform harmonic motion, with the same frequency and amplitude. For the harmonic wave, the wave equation is

$$E = E_0 \cos k\left(x - ct\right), \tag{13.3}$$

where E_0 is the wave amplitude (related to the energy carried by the wave), $k = \frac{2\pi}{\lambda} = 2\pi v$ is a wavenumber, λ is the wavelength and ν is the frequency. In addition, the period τ of oscillations is given as $\tau = \frac{2\pi}{\omega} = \frac{1}{\nu} = \lambda / \nu.$ Another form of 1-D wave equation is

$$E = E_0 \cos\left(kx - \omega t\right), \tag{13.4}$$

where $\omega = kc = \frac{2\pi c}{\lambda} = 2\pi v$ is angular frequency.

Wave Amplitude The amplitude of a wave may be constant (in which case the wave is a continuous wave) or may be modulated to vary with time and/or position. By introducing phase constant ϕ, Eq. 13.4 becomes

$$u(x,t) = A(x,t)\,\sin\,(kx - \omega t + \phi), \tag{13.5}$$

where A (x, t) is the amplitude of the wave, k is the wave number, ω is the frequency and ϕ is the phase.

Phase and Group Velocity It is known from kinematics that velocity is equal to the ratio of the relative distance travelled in a unit of time. How fast is the wave traveling depends on the phase velocity c, which is given by the relation $c = \lambda/\tau$. As the wave frequency $\nu = \frac{1}{\tau}$, then

$$c = \lambda \nu, \tag{13.6}$$

or in terms of k, $k = 2\pi/\lambda$ and the angular frequency $\omega = 2\pi/\tau$ for the phase velocity we obtain

$$c = \frac{\omega}{k}. \tag{13.7}$$

While the phase velocity gives the speed and the direction of phase propagation, group velocity indicates the speed and the direction of energy propagation. Later we will see that the atmospheric gravity waves are dispersive: the phase velocity is a function of the wavenumber, and thus, the phase and group velocities differ. In contrast, sound waves are not dispersive. The phase and group velocities of sound waves are equal, because they propagate in space with the speed of sound, regardless of their wavelength or frequency.

Standing Versus Traveling Waves

Standing (stationary) wave, represents a wave that remains in its constant position. This happens because the medium moves in the opposite direction of the wave. Another possible reason for the formation of a standing wave is the interference of two waves (with equal amplitude and frequency) that travel in opposite directions. Also, it is important to point out that standing waves are initiated when the boundary blocks further wave propagation, causing reflection and propagation of a new wave with the same characteristics but in the opposite direction. In these waves, the surfaces with a constant phase are static in relation to the Earth. But other waves that are not limited in a given space along the medium, i.e. surfaces move, so this type of wave model, which moves through the medium, is known as a traveling wave. A typical example of traveling waves are ocean waves.

13.3 Atmospheric Waves

13.3.1 History Studying Atmospheric Waves

The study of atmospheric motion in the atmosphere originated from Newton, who was dealing with sound waves (Saha 2008; Zdunkowski and Bott 2003). Despite such a long period in which various waves were being studied, the meteorologist's interest in them was especially noticeable for the past several decades. The atmosphere has a wave nature, with processes manifesting in a different spatial and temporal scale. For these reasons, the study of atmospheric waves is important for several reasons:

- Linearized equations have wave-type solutions which correspond with some observed types of waves in the atmosphere and could be mathematically solved.
- Since these waves are the solutions of our equations, they appear during the numerical integration.
- With some type of waves, huge amount of energy is transferred from one place of the atmosphere to another, similarly as electromagnetic waves transfer solar energy.
- The wave's amplitude under certain conditions rapidly increases, so some phenomena are generated, typical for non-linear cases.
- Gravity waves act as a triggering factor for instability, which initiates storms.
- In many circumstances, atmospheric stability prevails due to atmospheric adjustment. These are suitable conditions for the occurrence of waves.

13.3.2 Atmospheric Waves: Definition

Atmospheric waves are motions of air in the Earth's atmosphere which have different spatial (meters to thousands of kilometres) and temporal scales (minutes to weeks) (Volland 1988; Pedlosky 2003). There are many definitions of atmospheric wave. In general, an atmospheric wave is a periodic disturbance in the fields of meteorological variables (like pressure or geopotential height, temperature, wind or wind velocity), which may either propagate (traveling wave) or not (standing wave). Atmospheric waves can impact the wind, density, pressure or temperature fields and can be identified as fluctuations of these parameters. The waves are mainly formed in the troposphere, the lower layer of the atmosphere, and the stratosphere and propagate in the horizontal and vertical directions. Here, it is important to point out the difference between propagating and breaking waves. Propagating waves transport energy and momentum from their source regions across the atmosphere and can dissipate under certain conditions. Refractive waves are those whose amplitude reaches a critical level. As a result, sudden processes occur in which a large amount of energy is transformed into turbulent kinetic energy. As we pointed out above, atmospheric waves have a wide range of spatial and temporal dimensions, from large planetary waves (Rossby waves) to the smallest minute-sound waves.

13.3.3 Factors that Form a Wave

Wave-triggering factors are specific to certain types of waves. The most common factors are:

- Stability (expressed through density or entropy)
- Earth's rotation (expressed by the Coriolis parameter) or its change in latitude
- Daily variation of air heating by solar energy
- Electrodynamic (which are significant for the thermosphere)

Due to the diversity of waves, it is not possible to define waves in the atmosphere in a unique and precise way. However, they share the following characteristics: a certain quasiperiodic oscillations and ability to transmit "information" to large distances without proper displacement of particles of air. The term "information," among other things, refers to mass, momentum, energy, etc.

13.3.4 Basic Wave Properties and Classification

Waves in the atmosphere can be classified according to factors that give them characteristics or based on their geometric properties. So, there are sound, gravity, inertial, buoyancy, Rossby and other type of waves. According to their simplest geometry, the longitudinal, vertical-transferal and horizontal transferal waves differ. Longitudinal waves are characterized by the movement of particles of air along lines parallel to the direction of wave propagation. Longitudinal waves include sound waves (pressure perturbations, particle of displacement and particle velocity propagated in an elastic medium) and seismic P-waves (created by earthquakes and explosions). In vertical-transferal waves, the air particles move in a vertical plane and the wave extends horizontally. This type belongs to the most diverse wave of gravity type. And finally, the horizontal-transfer wave is characterized by the movement of air particles in the north-south direction and the wave extends west-east. The most important waves of this type are Rossby waves. There is also another classification that distinguishes the waves: free and forcing waves. Forcing waves continuously maintain their characteristics (phase velocity and wave number) by the forced mechanism. An example of such waves is thermal-induced waves due to daily warming from the Sun, the transfer of air over mountain obstacles, the flow around the growing convective cloud, etc. There are no such forcing mechanisms in the free waves. The identification of atmospheric waves is provided by direct observations and measurements or by applying the complex analysis of standard or special measurements (upper air sounding measurements, analysis of synoptic weather maps, meteorological bulletins and reports).

13.4 The Mathematical Concept of Atmospheric Waves

In Chap. 12, we dealt with the basic concept of atmospheric motion and various approximations to the equations of motion. However, it is very useful to look at simple but important types of wave motion in the atmosphere, which arises from this set of equation under different assumptions. For each case the appropriate equations are the momentum equations, the continuity equation and the thermodynamic equation as well as the total derivative of scalar quantity:

$$\frac{d\vartheta}{dt} = -\frac{1}{\rho}\nabla p - gk - 2\Omega x\vartheta + F \ (\text{Momentum equation}) \tag{13.8}$$

$$\frac{1}{\rho}\frac{d\rho}{dt} = -\nabla\cdot\vartheta \ (\text{Divergence form of the continuity equation}) \tag{13.9}$$

$$dQ = c_p dT - \alpha dp \ (\text{The first law of thermodynamics}) \tag{13.10}$$

$$\frac{d\psi}{dt} = \frac{\partial\psi}{\partial t} + \vartheta\cdot\nabla\psi \ (\text{The total derivative of scalar quantity}) \tag{13.11}$$

13.4.1 Atmospheric Sound Waves

Sound waves are pressure perturbation waves that travel through different media, such as the Earth's crust, water bodies and the atmosphere (Yang 2016). The atmosphere is spread in all possible directions by sound waves. Sound waves propagate at the speed of sound, which depends on the virtual air temperature.

The human ear hears a sound source of sufficient intensity if the frequencies are between 16 and 20000 Hz. Sound waves with a frequency of less than 16 Hz are called infrasound and those with more than 20000 Hz are called ultrasound. Figure 13.3 shows an example of sound waves in air as the pressure fluctuates in a

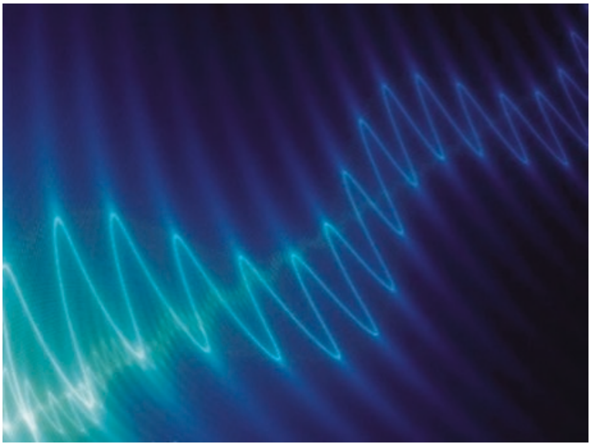

Fig. 13.3 A typical illustration of sound waves

sinusoidal variation. We will assume motion along x-axis and that the y and z components and gradients in these directions are zero. Coriolis and friction terms are also neglected, and the motion is considered adiabatic. With these approximations, the momentum equation (13.8) becomes

$$\frac{du}{dt} + \frac{1}{\rho}\frac{\partial p}{\partial x} = 0, \tag{13.12}$$

the continuity equation (13.9) transforms into

$$\frac{d\rho}{dt} + \rho\frac{\partial u}{\partial x} = 0, \tag{13.13}$$

and the first law of thermodynamics for adiabatic motion (Q = const.) follows as

$$p\rho^{-\gamma} = \text{const.}, \tag{13.14}$$

where $\gamma = \dfrac{C_P}{C_V}$ is the specific heat ratio under constant pressure and constant volume, respectively. Combining last two equations we obtain

$$\frac{dlnp}{dt} + \frac{\partial u}{\partial x} = 0. \tag{13.15}$$

The perturbation method allows variables to be written as the sum of a mean variable (given by bar) and a perturbed one (represented by prime):

$$u = \overline{u} + u'; p = \overline{p} + p'; \rho = \overline{\rho} + \rho'. \tag{13.15}$$

Here $p' \ll \overline{p}$, $\rho' \ll \overline{\rho}$, but u is not necessarily < $\overline{u}$ as the solutions are still valid if $\overline{u} = 0$.

Substituting in (13.12) and (13.15) where p and ρ appear, neglecting products of primed quantities, and if the differentials of mean quantities are zero, then using (13.11) we obtain

$$\left(\frac{\partial}{\partial t} + \overline{u}\frac{\partial}{\partial x}\right)u' + \frac{1}{\overline{\rho}}\frac{\partial p'}{\partial x} = 0 \tag{13.17}$$

$$\left(\frac{\partial}{\partial t} + \overline{u}\frac{\partial}{\partial x}\right)p' + \gamma\overline{p}\frac{\partial^2 u'}{\partial x^2} = 0. \tag{13.18}$$

Eliminating u' from these equations,

$$\left(\frac{\partial}{\partial t} + \overline{u}\frac{\partial}{\partial x}\right)^2 p' - \gamma\frac{\overline{p}}{\overline{\rho}}\frac{\partial^2 p'}{\partial x^2} = 0. \tag{13.19}$$

Equation (13.19) represents a wave equation with solutions:

$$\mathrm{p}' = \mathrm{R_e}\left\{\mathrm{A}\exp\mathrm{ik}\left(\mathrm{x} - \mathrm{ct}\right)\right\}, \tag{13.20}$$

with a constant A and

$$\mathrm{c} = \overline{\mathrm{u}} \pm \left(\gamma\frac{\overline{\mathrm{p}}}{\overline{\rho}}\right)^{1/2}. \tag{13.21}$$

The speed of sound relative to the flow $\overline{u}$ is therefore $\sqrt{\left(\gamma\frac{\overline{p}}{\overline{\rho}}\right)}$, where $\gamma = \frac{c_p}{c_v}$.

It is a characteristic quantity of the medium included in the system of hydrodynamic (gas-dynamic) equations and plays a significant role in the study of wave processes (Ćurić and Janc 2016). For these reasons, an accurate estimate of the speed of sound is essential for further adequate description of the formation and propagation of these waves in the medium. In the case when sound waves travel through an ideal gas, where there is no heat exchange due to faster expansion or

Table 13.1 Laplace's speed of sound as a function of virtual temperature

$T_v(^{\circ}C)$	−50	−30	−10	10	30
c (m/s)	300	313	325	338	349

compression of the gas, the longitudinal sound waves are adiabatic and the speed of sound is also called adiabatic speed of sound. Starting from the principles of dynamics and thermodynamics, we come to the speed of propagation of sound waves in the atmosphere, given by the following relations:

$$\mathrm{c} = \sqrt{\gamma \frac{\mathrm{p}}{\rho}} \ \mathrm{or}\ \mathrm{c} = \sqrt{\gamma RT} \ \mathrm{or}\ \mathrm{c} = \sqrt{\gamma R_s T_v}, \tag{13.22}$$

that is, speed of sound depends only on virtual temperature and does not depend on altitude. Based on French mathematician Pierre-Simon Laplace, speed c is called *Laplace's speed of sound*. Table 13.1 gives its values for different virtual temperatures. The effect of specific humidity on the speed of sound is very small. For example, if $q = 10g/kg$ and the speed of sound is 340 m/s for a dry air, then the speed of sound in humid air would be approximately 341 m/s, an increase of 0.3% over that in dry air.

So the speed of sound is a function of temperature only and to a lesser extent humidity. The explosion-induced waves propagate faster, and the faster the propagation speed, the stronger the explosion and the closer to the original location. Moving away from the explosion site, the velocity of propagation approaches Laplace. Because of this, sound waves propagate in all directions at the same rate in the dry isothermal atmosphere. The axes of the sound wave beam extend perpendicular to the sound source. The real atmosphere is usually not isothermal, so the sound waves do not spread everywhere at the same speed as the sound waves bend according to the law about breaking waves. Imagine that the two layers of air at different virtual temperatures touch each other sharply at a given boundary surface. Then, sound waves in all directions extend from the sound source located in the lower air layer. When passing over the boundary surface, the direction of propagation of the sound wave changes. If α and α' are the angles of incidence and refraction and c and c$'$ are speed of sound in the lower and upper layers, respectively, then according to Snelius's law of refraction,

$$\frac{sin\alpha}{sin\alpha'} = \frac{c}{c'} = \sqrt{\frac{T_v}{T_{v'}}} \tag{13.23}$$

If the lower layer of air is warmer, then by the previous formula the turning angle is less than the incidence angle, so the sound waves will break to normal. Conversely, if there is a temperature inversion on the ground floor, which means that the lower layer of air is cooler than the upper one, the sound waves will bend from the normal to the boundary surface. It follows from the above relation that the waves arriving normally on the boundary surface do not refract ($\alpha = \alpha' = 0$). For larger bump angles and refraction, it is increasing. For some value of the incidence angle, the wave extends over the boundary surface itself ($\alpha' = 0$). That breaking angle is

$$\sin \alpha_g = \sqrt{\frac{T_v}{T_{v'}}} \tag{13.24}$$

In the case of surface temperature inversion, when α > αg, the sound wave does not enter the warmer air and is already deflected according to the law of deflection so that the incident angle is equal to the deflection angle. In the atmosphere of a true boundary (discontinuous) surface, the direction of propagation of sound waves across various transition layers changes gradually. Figure 13.4 shows the direction of propagation of sound waves, when the temperature decreases with height (Fig. 13.4a) and when it increases with height (Fig. 13.4b). In the first case, the sound waves bend symmetrically from the source to the normal, so that two zones of silence are formed to the left and right of the waves that bounce to the Earth's surface. In the case of inversion, sound waves are bounced off normal, and there are no silence zones caused by the bending of the waves. Figure 13.5 shows the cases of sound wave deflection, in the presence of ground temperature inversion.

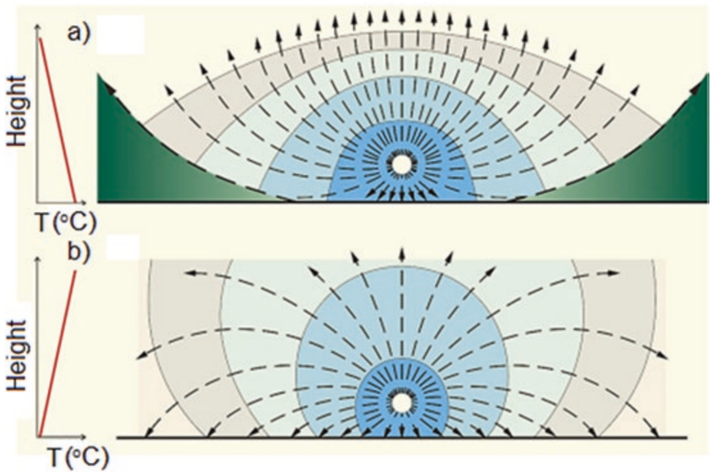

Fig. 13.4 Sound refraction due to change in temperature with altitude: (**a**) when the temperature decreases with altitude; (**b**) when the temperature increases with altitude. Areas in green are calm areas

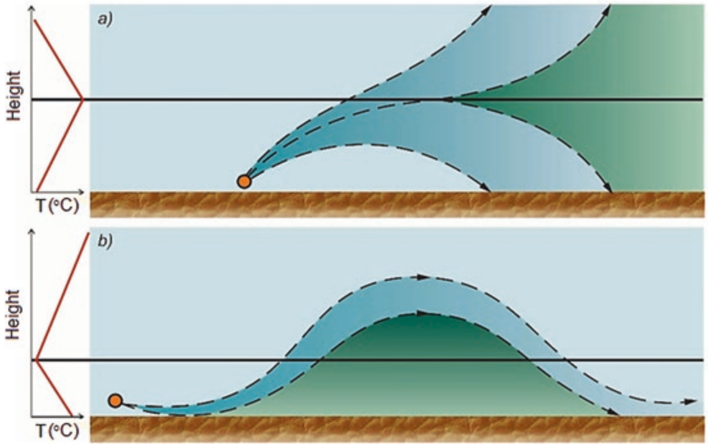

Fig. 13.5 Sound refraction in the case of temperature inversion: (**a**) a ground temperature inversion; (**b**) an *upper level thermal inversion*. The areas in green are the zones of windless weather conditions or calms

In the case of ground inversion, the sound waves in the inversion layer will bend from normal to take a convex shape. A sound wave that bounces about the upper inversion boundary at the greatest distance from the source and then returns downwards separates the audibility zone from the silence zone. Above the inversion layer, where the temperature drops with height, the sound beams bend upwards. The sound beam that will first enter the warmer air (after the boundary beam) represents the boundary of the zone of silence and audibility that is above the inversion layer. When the inversion is at height (Fig. 13.5b), the sound waves will first bend upwards from the source and then downwards, so that they return to the lower layer, which is warmer than the upper one, and then bend back to normal. The boundary of the audibility zone and the silence zone is determined by the boundary beam that touches the Earth's surface and then bends upwards.

13.4.2 Gravity Waves

Since sound waves are longitudinal, only one dimension needs to be considered obtaining the sound wave solution. Other waves exist where the oscillation transfers to the direction of propagation. Establishing equal pressure between air parcel and the environment leads to emission of sound waves, while establishing of equilibrium in air density leads to *gravity waves* (Nappo 2013). The existence of gravity waves is illustrated once we considere static gravitational instability and the buoyancy effect of a parcel displaced vertically from its equilibrium level with the frequency of oscillation (the Brunt–Väisälä frequency). To get a satisfactory solution to the basic equations for gravity waves, we consider a 2-D flow and arrive at the following approximations:

(i). A 2-D motion in only (x, z) frame.
(ii). Assume that the horizontal scale of the wave is small so that the Coriolis term may be neglected.
(iii). Ignore the friction term.
(iv). Assume also that the atmosphere is unperturbed at rest. Under such assumptions, the (x, z) equations of motion (13.8) are

$$\frac{du}{dt}+\frac{1}{\rho}\frac{\partial p}{\partial x}=0 \tag{13.25}$$

$$\frac{dw}{dt}+\frac{1}{\rho}\frac{\partial p}{\partial z}+g=0 \tag{13.26}$$

The continuity equation (13.9) becomes

$$\frac{1}{\rho}\frac{d\rho}{dt}+\frac{\partial u}{\partial x}+\frac{\partial w}{\partial z}=0. \tag{13.27}$$

And the first law of thermodynamics for adiabatic motion means that the potential temperature θ remains constant, that is,

$$\frac{dln\theta}{dt}=0. \tag{13.28}$$

From the equation for the potential temperature expressed in logarithmic form in terms of ρ and p such that

$$\frac{dln\theta}{dt}=\frac{1}{\gamma}\frac{dlnp}{dt}-\frac{dln\rho}{dt}. \tag{13.29}$$

For the unperturbed atmosphere, neither θ, p nor ρ vary in the horizontal. When we apply perturbation method in the above equations, for the derivation of sound waves, it becomes

$$u=u';w=w';p=\overline{p}+p';\rho=\overline{\rho}+\rho'.$$

Then using the hydrostatic equation, the equation of state, where H is the scale height, also neglecting the products of perturbed quantities, the four equations (13.25), (13.26), (13.27), (13.28) and (13.29) for the unknown $u', w', \rho'/\bar{\rho}$ and $p'/\bar{p}$ become

$$\frac{\partial u'}{\partial t} + gH\frac{\partial}{\partial x}\left(\frac{p'}{\bar{p}}\right) = 0 \tag{13.30}$$

$$\frac{\partial w'}{\partial t} + g\left(\frac{\rho'}{\bar{\rho}}\right) + gH\frac{\partial}{\partial z}\left(\frac{p'}{\bar{p}}\right) - g\left(\frac{p'}{\bar{p}}\right) = 0 \tag{13.31}$$

$$-\frac{N_B^2}{g}w' + \frac{\partial}{\partial t}\left(\frac{\rho'}{\bar{\rho}}\right) - \frac{1}{\gamma}\frac{\partial}{\partial t}\left(\frac{p'}{\bar{p}}\right) = 0, \tag{13.32}$$

where $\frac{N_B^2}{g}$ is the vertical stability parameter written for $\partial ln\theta/\partial z$. In solving these equations, we assume an isothermal atmosphere in which the scale height H is a constant. We will look for the solution of the equations (13.30), (13.1) and (13.32) for each unknown perturbed variables in the form exp (αz) exp $i(\omega t + kx + mz)$. On equating the imaginary parts, we find that $= \frac{1}{2H}$. Solutions with the first condition $m = 0$ have no phase variation in the vertical and are known as *external gravity waves* (whose maximum amplitudes are at the boundary of fluid). When $m \neq 0$, we have internal gravity waves (oscillation occurs in a stratified fluid and develop on the interfaces of density discontinuity). Substituting $\alpha = \frac{1}{2H}$ into the equations, the following dispersion relation results

$$m^2 = k^2\left(\frac{N_B^2}{\omega^2} - 1\right) + \frac{\left(\omega^2 - \omega_a^2\right)}{c^2}, \tag{13.33}$$

where c is the velocity of sound:

$$N_B^2 = \frac{(\gamma - 1)g}{\gamma H} \tag{13.34}$$

is the square of the Brunt–Väisälä frequency for the isothermal atmosphere and

$$\omega_a = \frac{1}{2}\left(\frac{\gamma g}{H}\right)^{1/2} = \frac{c}{2H} \tag{13.35}$$

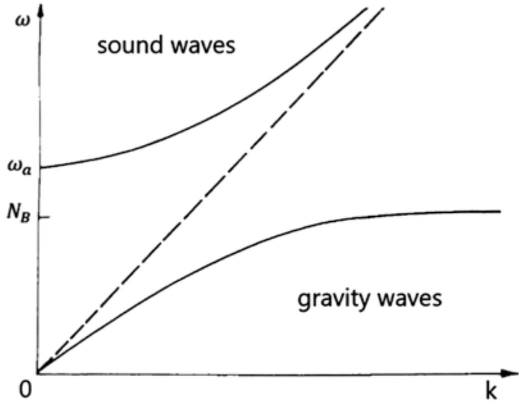

Fig. 13.6 Dispersion curves for gravity waves

is known as the acoustic cut-off frequency. The dispersion relation (13.33) is illustrated in (Fig. 13.6), where the curves of m = 0 are plotted on a $\omega - k$ diagram. Two regions where m^2 is positive are apparent, the higher frequency region ($\omega > \omega_a$) when $k = 0$ describes acoustic waves and the lower frequency region describes the gravity waves.

If the assumption that the atmosphere is isothermal is omitted, (13.33) may still be used as a reasonable approximation, with $N_B < \omega_a$ as the Brunt– Väisälä frequency for the atmosphere, which is considered, provide $N_B < \omega_a$ as is normally the case throughout the atmosphere. If the atmosphere is in a state of uniform zonal flow $\bar{u}$, (13.33) still holds provided ω is replaced by $\omega + \bar{u}k$, the frequency which would be seen by an observer moving with a basic flow and known as *intrinsic frequency*. Equation (13.33) applies to waves having a wide range of wavelength, frequency and velocity. For gravity waves of horizontal wavelength of the order of a few kilometres, the first term is much larger than the second. In this case, allowing also for the presence of a uniform zonal wind $\bar{u}$, the dispersion relation (13.33) may be written as

$$\frac{\left(\omega - \bar{u}k\right)^2}{N_B^2} + \frac{k^2}{m^2 + k^2}. \tag{13.36}$$

The characteristic gravitational waves that are often observed on the leeward sides of mountains are called lee waves; in meteorology, they are known as atmospheric stationary waves. Atmospheric internal gravity waves will be stationary with respect to the mountain and the horizontal component of phase velocity relative to the surface will be zero, in which case, for $k \ll m$ from (13.36), we have

$$m = \frac{N_B}{\bar{u}}. \tag{13.37}$$

Fig. 13.7 Rotor clouds

The first theory of stationary orographic waves originated from Lyre (1943) and is therefore called Lyra's theory of stationary mountain waves. Atmospheric internal gravitational waves are a typical example of orographic waves. They occur when the air is forced to flow over a mountain in a stable and continuously stratified fluid as it moves downstream from the topography. Orographic clouds associate with strong vertical motions, that usually develop in the form of turbulent eddies on the leeward side of mountain are known as rotor clouds (see Fig. 13.7).

Rotors are turbulent rotating winds that form in the lower part of the atmosphere under the crests of mountain waves and belong to the family of so-called altocumulus-lenticularis. On the Earth's surface, the wind is in the opposite direction from the gradient wind. Since the energy source is at the near surface layer, the group velocity of rotors travels upwards, while the phase velocity downwards. Consequently, the lines of constant phase in the stationary waves tilt with height backward relative to the mean flow; although in practice, because the mean wind varies with height, the situation is rarely as simple as this.

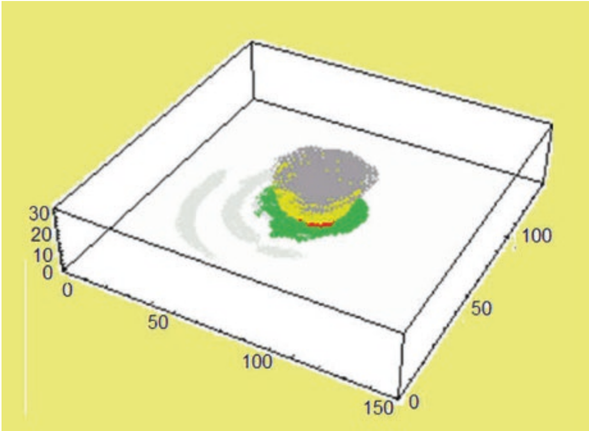

Fig. 13.8 A thermally induced inertial-gravity waves in a simulated 3-D convective cloud

13.4.3 Inertial-Gravity Waves

Gravity waves in a rotating system with a large wavelength are referred as *inertial-gravity waves*. They usually occur when a statically stable flow is also inertial stable. Fig. 13.8 shows an example of inertial gravity waves generated in a simulated storm using a 3-D cloud resolving (Spiridonov et al. 2020). Due to the presence of the Earth's rotation, for the physical description of these waves, it is necessary to include the Coriolis parameter in the basic equations. For simplicity, we will use the incompressible continuity equation.

Therefore, the linearized governing equations are:

$$\frac{\partial u'}{\partial t} = -\frac{1}{\bar{\rho}}\frac{\partial p'}{\partial x} + fv' \tag{13.38}$$

$$\frac{\partial v'}{\partial t} = -\frac{1}{\bar{\rho}}\frac{\partial p'}{\partial y} - fu' \tag{13.39}$$

$$\frac{\partial w'}{\partial t} = -\frac{1}{\bar{\rho}}\frac{\partial p'}{\partial z} - \frac{\rho'}{\bar{\rho}}g \tag{13.40}$$

$$\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} \tag{13.41}$$

$$\rho' = \frac{\bar{\rho}}{g}N^2\Delta z \tag{13.42}$$

Combining (13.40) with (13.42) to eliminate ρ' yields

$$\frac{\partial^2 w'}{\partial t^2} = -\frac{1}{\bar{\rho}}\frac{\partial^2 p'}{\partial z \partial t} - N^2 w'. \tag{13.43}$$

Equations (13.40), (13.41), (13.42) and (13.43) are the governing equations for inertial-gravity waves. Assuming the standard sinusoidal solutions for u', v', w' and p' in form of $A_i e^{i(kx + ly + mz - \omega t)}$ with the corresponding amplitude (A_1, A_2, A_3 and A_4) and substituting into (13.38), (13.39), (13.40) and (13.41) after some replacements, the resulting dispersion relation for inertial-gravity waves is

$$\omega^2 = \frac{\left(f^2 m^2 + N^2 K_H^2\right)}{K^2}. \tag{13.44}$$

In case the effects of Earth's rotation are ignored ($f = 0$), then the dispersion relation becomes the same as for pure internal waves. Since inertial-gravity waves have long wavelengths to be affected by the Earth's rotation, we can assume that they are in hydrostatic balance. This implies that $m \gg K_H$. Therefore, the dispersion relation for inertial-gravity waves could be written as

$$\omega^2 \cong f^2 + \frac{N^2 K_H^2}{m^2}. \tag{13.45}$$

13.4.4 Rossby Waves

From the wave motions which form into homogeneous fluid, for large-scale atmospheric processes, the most important are the Rossby waves (Ćurić and Janc 2002).

They represent the very large-scale wave features which are observed on the flow of the planetary scale (Fig. 13.9). For large-scale meteorological processes, Rossby waves are the most significant of all waves. They are called Rossby waves after the famous Swedish meteorologist Gustav Rossby, who was the first to bring them in connection with meteorological phenomena. They are conditioned by a change in the Coriolis force with latitude, which means that they are directly related to the rotation of the Earth. In a -viscous barotropic fluid $\rho = \rho\ (p)$ of constant thickness in which the divergence of the horizontal velocity vector is equal to zero, the Rosby wave is a motion in which absolute vorticity is conserved. The appropriate momentum equations (13.8) are:

$$\frac{du}{dt} + \frac{1}{\rho}\frac{\partial p}{\partial x} - fv = 0 \tag{13.46}$$

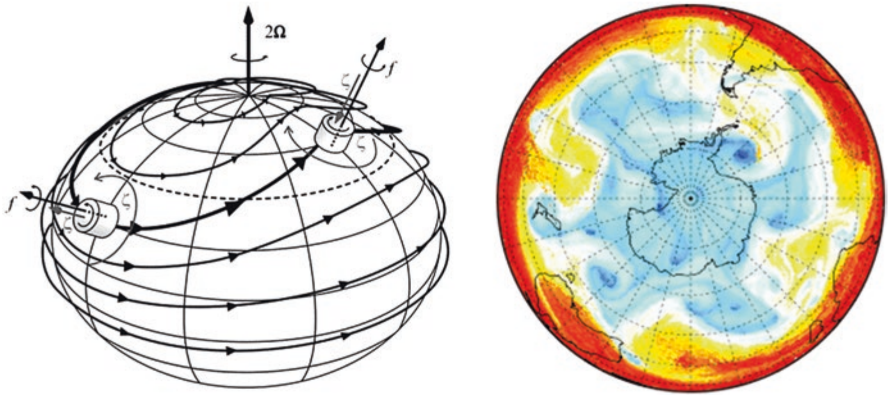

Fig. 13.9 Rossby waves. Scheme of "air particle" reaction to meridional displacement. Zonal movements when distorted to the south receive cyclone vorticity (to maintain absolute vorticity). Moving the air parcels northwards, they gain anticyclone vorticity (left panel). In Fig. 13.9, a graphical illustration of the planetary Rossby waves is represented through global distribution of geopotential height and potential vorticity surface (gpm) in the Southern Hemisphere (right panel)

$$\frac{dv}{dt} + \frac{1}{\rho}\frac{\partial p}{\partial y} + fu = 0 \tag{13.47}$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0. \tag{13.48}$$

Operating on (13.46) with $\partial/\partial y$ and (13.47) with $\partial/\partial x$ and subtracting

$$\frac{d}{dt}\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right) + f\left(\frac{\partial v}{\partial y} + \frac{\partial u}{\partial x}\right) + v\frac{\partial f}{\partial y}. \tag{13.49}$$

The quantity $\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$ is the vertical component of relative vorticity ζ; it may be considered as a vector equal to twice the local angular velocity of fluid elements. The second term in the last equation is zero by (13.48); that means horizontal wind divergence is zero. Because the Coriolis parameter f (vertical component of Earth's vorticity relative to a fixed system) is

$$\frac{d(\zeta + f)}{dt} = 0. \tag{13.50}$$

The quantity $\xi_a = \zeta + f$ is known as absolute vorticity (vorticity due to the rotation of the fluid itself combined with that due to the Earth's rotation). Equation (13.50) shows that under the assumption of a non-divergent and frictionless flow, absolute vorticity is conserved. In order to solve this equation, we assume a linear relation between f and y. It supposes that $f = f_0 + \beta y$, where $\beta = df/dy$ is the gradient of planetary vorticity at the initial latitude, known as the *β –plane approximation*. If we assume a uniform zonal flow $\bar{u}$ in the unperturbed case and introduce perturbations, that is $u = \bar{u} + u', v = v'$, (13.50) becomes

$$\left(\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x}\right)\left(\frac{\partial v'}{\partial x} - \frac{\partial u'}{\partial y}\right) + \beta v' = 0. \tag{13.51}$$

Since the flow is non-divergent in the horizontal, *a stream function* ψ can be introduced such that using (13.48) is automatically satisfied, that is,

$$u' = -\frac{\partial \psi}{\partial y}, v' = -\frac{\partial \psi}{\partial x}. \tag{13.52}$$

Substituting in (13.51),

$$\left(\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x}\right)\nabla^2\psi + \beta\frac{\partial \psi}{\partial x} = 0. \tag{13.53}$$

For wave solutions $\psi = R_e\{\psi_0 \exp i(\omega t + kx + ly)\}$ to be possible, the dispersion relation

$$c = -\frac{\omega}{k} = \bar{u} - \frac{\beta}{k^2 + l^2} \tag{13.54}$$

must be satisfied. The velocity relative to the zonal flow is $c - \bar{u}$, where c is the phase velocity in the x direction. Rossby waves, therefore, drift to the west relative to the basic flow, at typical speeds of a few metres per second. Note that the phase speed of the waves increases with wavelength. In a more complex fluid, such as the baroclinic atmosphere, Rossby waves are movements where the potential vorticity is conserved, and it is formed due to the existence of an isentropic gradient of potential vorticity.

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Chapter 14
Atmospheric Boundary Layer (ABL)

Atmospheric boundary layer (ABL) also known as *planetary boundary layer* (PBL) also determined as the region of the lower levels of the troposphere (Sanchez-Lavega 2011). Before studying in detail the influence of the Earth's surface on the main physical processes in the atmosphere (e.g. heat, moisture and momentum exchange through turbulence, mixing and convection), we will briefly focus on the historical development of ABL.

14.1 ABL Historical Overview

The origin of the aerodynamic laminar boundary layer theory, starting with a seminal paper of Ludwig Prandtl (1875–1953), a German engineer, who has developed boundary layer theory in 1904. One year later Vagn Walfrid Ekman (1874–1954), Swedish oceanographer, developed his theory of laminar Ekman layer and wind spiral in ABL; Osborne Reynolds (1842–1912), Ireland mathematician, developed fluid dynamics and Reynolds number; Andrey Nikolaevich Kolmogorov (1903–1987), Russian mathematician, developed similarity theory of turbulence; and from 1910 to 1940, Sir Geoffrey Ingram Taylor develops basic methods for examining and understanding turbulent mixing. During this period also mixing-length theory and eddy diffusivity approach have been developed by von Karman, Prandtl and Lettau. Discovery of similarity theory of turbulence by Andrey Nikolayevich Kolmogorov (1903–1987), Russian mathematician, advanced the understanding of turbulent processes in that time. Buoyancy effects on surface layer were studied by Monin and Obuhkov from 1950 to 1960. The next decade from 1960 to 1970 was the Golden Age of Boundary Layer Meteorology; accurate observations of a variety of boundary layer types, including convective, stable, and trade cumulus, started and verification/calibration of surface similarity theory. The period from 1970s–1980s is marked with development of resolved three-dimensional (3D) computer modelling of boundary layer (BL) turbulence (large eddy simulation or LES) as well as the application of

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higher-order turbulence closure theory. Ten years later major field efforts were given in stratocumulus-topped boundary layers, land surface vegetation parameterization, and mesoscale modelling. The rapid technology developments from 1990 to 2000 allowed utilization of new surface remote sensing tools (e.g. LIDARs, cloud radars) and extensive space-based coverage of surface characteristics. Advanced observation and modelling studies significantly improve the knowledge and understanding of boundary layer, the physical processes, deep convection, land surface processes, and land-ocean interactions (Oke 2002). Coupled ocean-atmosphere-ice-biosphere models created new requirements for boundary layer parameterizations, e.g. better treatment of surface wind stress, vegetated surfaces, BL clouds, and aerosols (Wells 2012). In order to successfully model the air flow around buildings in urban areas as well as air pollution in complex terrains, an even more precise boundary layer simulation is needed. Ensemble data assimilation enables better use of near surface and boundary layer data over land surfaces.

14.2 ABL Definition and Basic Characteristics

Atmospheric boundary layer (ABL) is the bottom layer of the troposphere that is in contact with the Earth's surface (Baklanov et al. 2007; Garratt 1994; Schlichting and Gersten 2017; Lee 2018). ABL is that region of the atmosphere in which the surface influences fluctuations of the physical quantities (e.g. temperature, moisture, wind). It is the portion of the troposphere that is directly influenced by the Earth's surface and responds to combined action of mechanical and thermal forcing, in the order of 1-h time scale. The basic feature of the ABL is the presence of frequent turbulence generated by buoyancy, unstable stratification, surface warmer than air, and/or mechanical mixing due to wind shear. It is one of the main mechanisms for generating positive buoyancy which is radiative heating at the Earth's surface. This energy is transmitted upwards to the atmosphere by conduction and convection. Convective transfer usually takes the form of convective plumes. Dominant processes in the PBL are as follows:

- Radiative forcing
- Large diurnal temperature variations (relative to the free troposphere)
- Subscale atmospheric processes
- Transport and exchange of heat, moisture, momentum
- Turbulence
- Mixing
- Viscosity-surface friction

In a free atmosphere, the geostrophic wind blows independently of surface conditions. The geostrophic wind is influenced by the Earth's rotation and is the driving force for all winds in the underlying layers. The free atmosphere starts at around 1000 m depending on the thickness of the underlying atmospheric boundary layer (varies between a few 100 m and 3000 m). The atmospheric boundary layer is divided into two sublayers:

1. the surface layer; and
2. the Ekman layer

The ground-level boundary layer (Prandlt layer) occupies about 10% of the atmospheric boundary layer. Below the ground-level boundary layer, there is no wind. The different wind directions of the geostrophic winds and the ground-level winds lead to a rotation of the wind direction within the Ekman layer. Turbulence by itself cannot transfer heat (or momentum, moisture) across the interface between the atmosphere/ocean and atmosphere/earth. There are some other mechanisms responsible for the transport of these variables in the lowest few millimetres of the atmosphere (laminar boundary layer). Molecular conduction transfers heat between the surface and the lowest part of the atmosphere. Once in the air, turbulence takes over to transport heat to greater depths in the atmosphere. A similar argument applies for the transfer of momentum and moisture in the atmosphere. When all elements of fluid are moving along the direction of the fluid's mean motion, the flow is laminar, and the internal friction in the fluid will be due only to molecular viscosity. When individual elements are moving irregularly compared with the mean motion, the flow is turbulent. Reynolds number is a non-dimensional quantity which determines whether the shared flow is turbulent or laminar (Fedorovich et al. 2004).

$$R_e = LV / \nu \tag{14.1}$$

where L is the typical length scale of the motion, V is the fluid velocity, and ν is the kinematic viscosity. For larger values of R_e = 6000, turbulence occurs. The main sources of turbulence are:

- Thermal forcing (thermals: buoyant eddies forced by solar heating of the surface)
- Vertical wind shear (due to frictional drag by the surface on geostrophic flow aloft).

Other processes or phenomena shown on Fig. 14.1, which also occur in the PBL, represent:

(i). *Horizontal wind shear* (due to flow of wind around obstacles: trees, mountains, islands).
(ii). *Gravity waves* (gravity).
(iii). *Spiralling winds* (related to neutral boundary layers and Ekman wind spiral).
(iv). *Katabatic winds* (downsloping or gravity winds, which blow down a slope as the result of gravity (Markowski and Richardson 2010). It occurs at night-time radiative cooling.
(v). *Nocturnal jet* (strong radiation cooling during night in a few hundred meters this effectively decouples the higher flow from the friction associated with the boundary layer.
(vi). *Clouds* (fair weather cumulus clouds (whose roots are in the BL), trade cumulus (which may rain), stratocumulus clouds (which may rain) or fog.

The basic difference between the PBL and the free atmosphere is listed in Table 14.1.

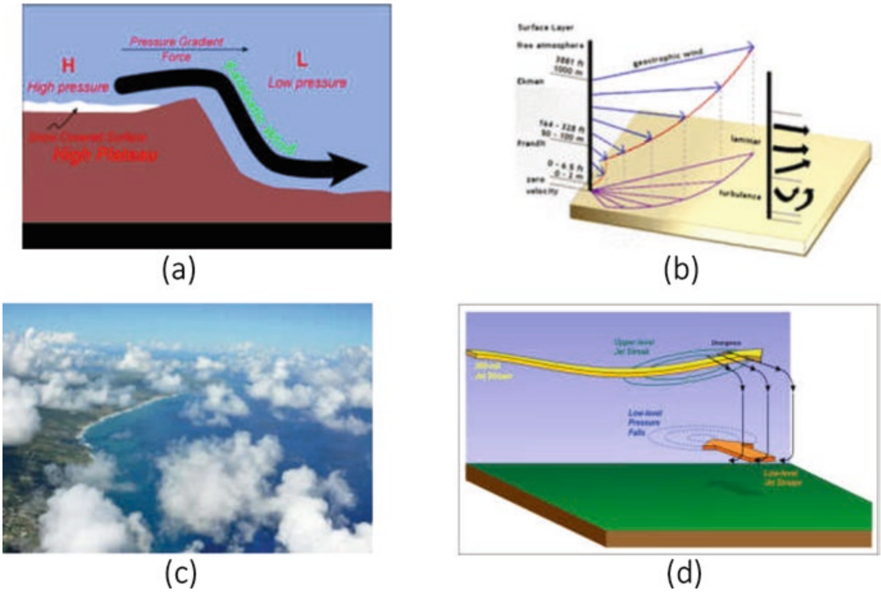


Fig. 14.1 Graphical illustration of PBL phenomena. (**a**) Katabatic winds; (**b**) spiralling winds; (**c**) trade cumulus clouds; (**d**) low-level jet stream

Table 14.1 Boundary layer vs free atmosphere

Property	Boundary layer	Free atmosphere
Turbulence	Almost continuously turbulent over its whole depth large horizontal extent	Turbulence in convective cloud and sporadic cat in thin layers of large horizontal extent
Friction	Strong drag against the Earth's surface. Large energy dissipation	Small viscous dissipation
Dispersion	Rapid turbulent mixing in the vertical and horizontal	Small molecular diffusion. Often rapid horizontal transport by mean wind
Winds	Near logarithmic wind speed profile in the surface. Sub-geostrophic, cross-isobaric flow common	Winds nearly geostrophic
Vertical transport	Turbulence dominates	Mean wind and cumulus scale dominate
Thickness	Varies between 100 m and 3 km in time and space. Diurnal oscillations over land	Fewer variables from 8 to 18 km. Slow time variations

14.2.1 ABL Significance

The lowest portion of the atmosphere is the most important for the human life and activities. Emission of various gases, source, and sink region of many trace gases (including water vapour, CO_2, ozone, methane) and dusts/pollutants occur in this layer. It is also significant for the air quality, especially urban air quality and pollutants emission, transport, accumulation, and dispersion. In the near surface layer, the main physical processes and fluxes take place. It is estimated that about 50% of the atmosphere's kinetic energy is dissipated in the boundary layer. It is important for local forecasting, cloud chemistry processes, which are very important for climate.

14.3 ABL Structure

As it is shown on Fig. 14.2, atmospheric boundary layer contains the following main sub-layers:

- Surface layer
- Mixing layer
- Stable boundary layer
- Residual layer
- Free atmosphere

In general, PBL is a layer where the air movement is mostly turbulent. The fluxes of momentum, heat, or matter are carried by turbulent motions on a scale of the order of the depth of the boundary layer or less (see Fedorovich et al. 2004). The surface effects (friction, cooling, heating, or moistening) are felt on time scales < 1

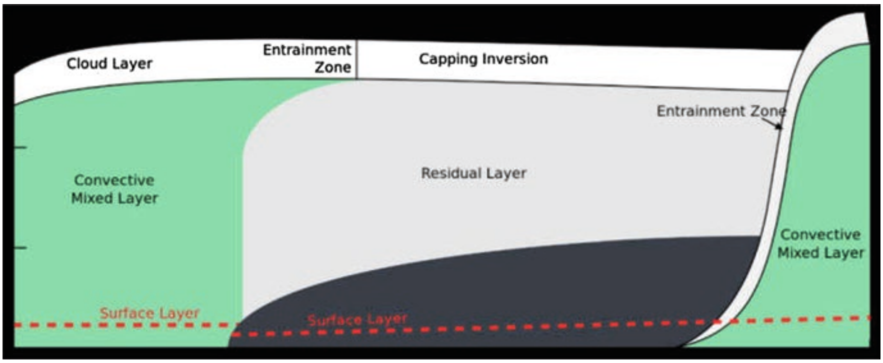


Fig. 14.2 Atmospheric boundary layer (ABL structure). Credit: NikNaks [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)]

day. The main composition of ABL layer is the atmospheric gases (N_2, O_2, water vapour, aerosol particles, clouds-condensed water).

14.3.1 Surface Layer

This layer is found at the bottom and affects about 1/10 of the ABL. This narrow area shown on Fig. 14.3 is most influenced by surface properties like heat fluxes, moisture, and mass. Intense small-scale turbulence generated by the surface is roughness and effects of convection. Earth rotation is negligible. Roughness layer is below the surface layer. Flow is highly irregular strongly affected by the nature of the individual roughness features (e.g. grass, trees, woods, buildings). Surface roughness determines to a certain extent the amount of turbulence production, the surface stress, and the shape of the wind profile.

14.3.2 Mixing Layer

Mixing layer (see Fig. 14.4a) is associated with unstable atmospheric conditions with strong upward heat flux from the surface of the Earth and low wind speeds. Generally, the planetary boundary layer is associated with buoyant thermals (unstable parcels of air) rising from the surface layer.

During these unstable conditions, which usually occur during daytime, the planetary boundary layer is called the mixed layer.

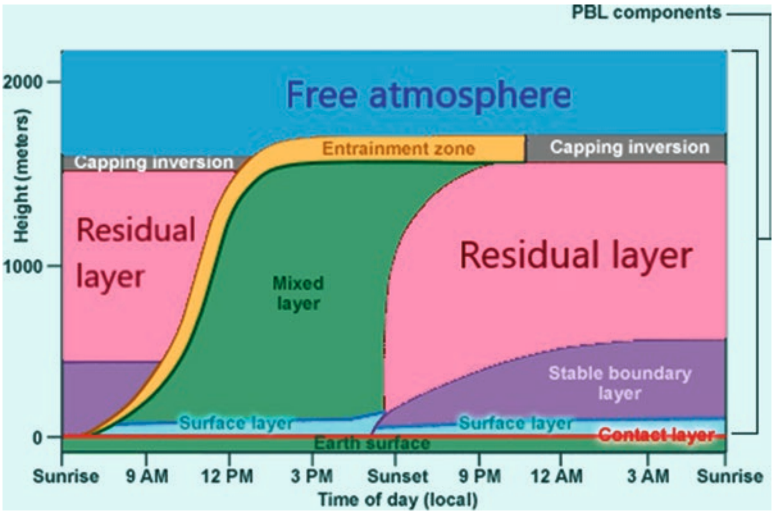
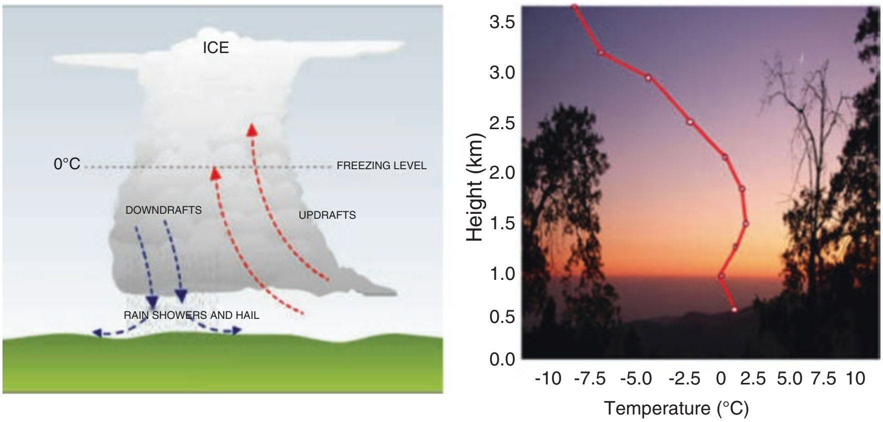


Fig. 14.3 Surface layer (light blue patterns); residual layer (two shaded areas with pink color)



(a) (b)

Fig. 14.4 (**a**) Mixing layer; (**b**) stable boundary layer

14.3.3 Residual Layer

It is the part of the atmosphere where mixing still takes place as a result of air flow, although heat fluxes from the surface of the Earth are small. The top of the residual layer was very near the elevation of an inversion in yesterday's sounding, as one might expect. Below that inversion, the atmosphere yesterday afternoon was relatively well mixed.

14.3.4 Stable Boundary Layer

This layer shown on Fig. 14.4b is characterized with radiation cooling of the air just above the surface, which tends to create a low-level inversion with relatively "stable" conditions. The depth where temperatures increase from the surface of the Earth up to the inversion level is called the stable boundary layer and ranges from about 100 m to 500 m deep. Radiation cooling is responsible for forming nocturnal stable boundary layers. It is also associated with formation of dew, frost, and fog if the humidity is sufficiently high.

14.3.5 Free Atmosphere

The free atmosphere is that portion of the Earth's atmosphere, which lies above the planetary boundary layer. In a free atmosphere the effect of the Earth's surface friction on the air motion is negligible. The air is usually treated (dynamically) as an ideal gas. The base of the free atmosphere is usually taken as the geostrophic wind level due to the balance between horizontal pressure gradient force and Earth's gravitation (described in Chapter 12 related to atmospheric motion).

14.4 Factors Influence on ABL Structure

The structure of the atmospheric boundary layer is influenced by:

- The underlying surface
- The stability of the PBL

Stability influences the structure of turbulence. In a stable stratified boundary layer (e.g. during night-time over land), the turbulence produced by shear is suppressed by the stratification resulting in a weak exchange and a weak coupling with the surface.

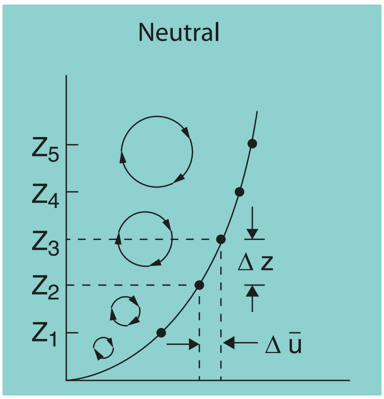
14.4.1 The Neutral PBL

The neutral PBL is graphically illustrated on Fig. 14.5. It is defined when the surface buoyancy flux is exactly zero (i.e. no buoyancy forcing for turbulence). More precisely we regard a boundary layer as neutral if shear production of turbulent kinetic energy is much larger than buoyant production. This condition rarely occurs outdoors particularly over land. After sunset, the ground becomes colder than the air above it, the surface buoyancy flux becomes negative, and the PBL becomes “stable”. The definition of stable or unstable clear PBL is determined by the sign of the surface buoyancy flux, not by the potential temperature gradient within the PBL. A well-mixed layer where $\partial\Theta/\partial z \sim 0$ is usually a neutral PBL because such a well-mixed layer is likely to be maintained by strong surface heating).

14.4.2 The Ekman Wind Spiral

Wind profile in the neutral boundary layer is called the Ekman wind spiral (Ćurić 2000). An important factor in the atmospheric boundary layer is that winds are turbulent. This acts as a frictional force in the momentum balance that affects the wind

Fig. 14.5 Neutral layer



profile. With suitable assumptions, the wind profile can be estimated through the entire boundary layer analytically. It is rare that this is possible with the equations of motion because of their non-linearity. In 1905, Walfrid Ekman, using some assumptions, derived an expression for the velocity profile in a neutral boundary layer. The resulting wind profile is a spiral, called the Ekman spiral (see Fig. 14.6).

14.4.3 Unstable Stratified Atmospheric Boundary Layer

In an unstable stratified PBL (see Fig. 14.7), during daytime over land with an upward heat flux from the surface, the turbulence production is enhanced, and the exchange is intensified resulting in a more uniform distribution of momentum, potential temperature, and specific humidity.

Based on the temperature gradient using virtual potential temperature as the conservative variable for moist atmosphere (Fig. 14.8), PBL is classified as:

1. Stable layer $\dfrac{\partial \bar{\theta}_v}{\partial z} > 0$
2. Unstable layer $\dfrac{\partial \bar{\theta}_v}{\partial z} < 0$
3. Neutral layer $\dfrac{\partial \bar{\theta}_v}{\partial z} = 0$

In reality, the PBL is more often thermally stratified (non-uniform temperatures and thus buoyancy mixing are now important in addition to mechanical).

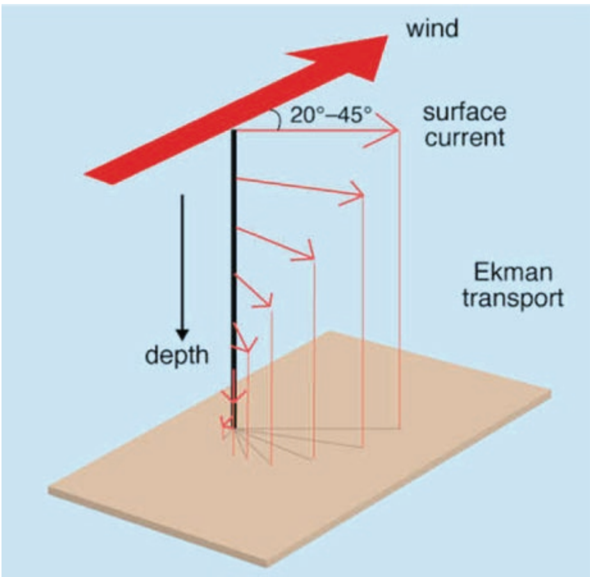


Fig. 14.6 The Ekman spiral

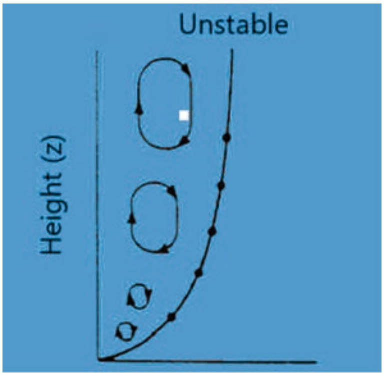


Fig. 14.7 Unstable stratified PBL

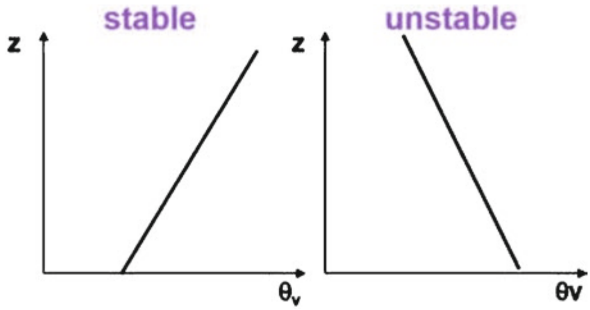


Fig. 14.8 Stable vs unstable layer

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Chapter 15
General Circulation of the Atmosphere

15.1 General Atmospheric Circulation Definition

There are two basic factors that determine the general circulation of the Earth's atmosphere: the latitudinal change of energy received by the Earth-atmosphere system from the Sun and the corresponding global distribution of angular momentum of the atmosphere (Satoh 2014). Air circulation over the Earth is mainly due to non-equal heating of the Earth's surface. The Earth's energy balance involves incoming solar radiation (S) and outgoing thermal infrared radiation (I), which together make up the net radiation (R_n).

$$R_n = S\left(1 - \alpha_p\right) + \left(I \downarrow - I -\right) \tag{15.1}$$

where α_p is planetary albedo ~0.3, I↓ is downward infrared radiation, and I↑ represents upward infrared radiation. The maximum transport of both properties occurs about the 40 degrees in each hemisphere. The total energy transport maximum is about 6 x 10^{15} W, of which ocean currents account for about 6 x 10^{15} W. The global atmospheric circulation, pressure distribution, and wind patterns play an integral role in the heat balance of the Earth and the creation of global ocean currents (Randall 2000). At the global atmospheric circulation, there is a transport of warm air from lower latitudes to higher latitudes and cold air from high latitudes to low latitudes. This heat exchange keeps the regions in low latitudes warmer, where there is an excess of total flow of energy throughout the year as a result of continuous heating, and high latitudes colder as a result of continuous cool due to the total energy loss.

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15.2 Conceptual Model of the General Atmospheric Circulation

The first model for the general circulation of the atmosphere was introduced by Edmond Halley in 1685. It is thermally direct meridional cell with maximum heating in low latitudes. According to the pattern, air rises and flows towards the poles and sinks as it cools. George Hadley in 1735 improved this concept by taking into consideration the Earth's rotation in flow deflection as the result of the Coriolis force, to the right at the Northern Hemisphere, but overlooked the westerly wind belt. A more detail picture about the global atmospheric circulation pattern was developed by William Ferrel, who proposed a three-cell model, where the role of the angular momentum conservation in westerlies was significant. His consideration was like that of C.-G. Rossby in 1941, with two direct cells bounding. The indirect one found in middle latitudes he prescribed rather to easterlies rather than westerlies. The solution appeared with the problem of strong westerlies was resolved in the 1950s when the role of a short wave (horizontal eddies) in poleward energy transfer was recognized. Because the patterns of circulation between the equator and 30° N and 30° S cross the northern and southern latitude have a similar appearance to the first single-cell model used by George Hadley in 1735, it is called "Hadley Cell Circulation".

15.3 Three-Cell Model of Circulation

The circulation of wind in the atmosphere is driven by the rotation of the Earth and the incoming energy from the Sun. Wind circulates in each hemisphere in three distinct cells which help transport energy and heat from the equator to the poles. The winds are driven by the energy from the Sun at the surface as warm air rises and colder air sinks. The three-cell circulation model shown on Fig. 15.1 describes the

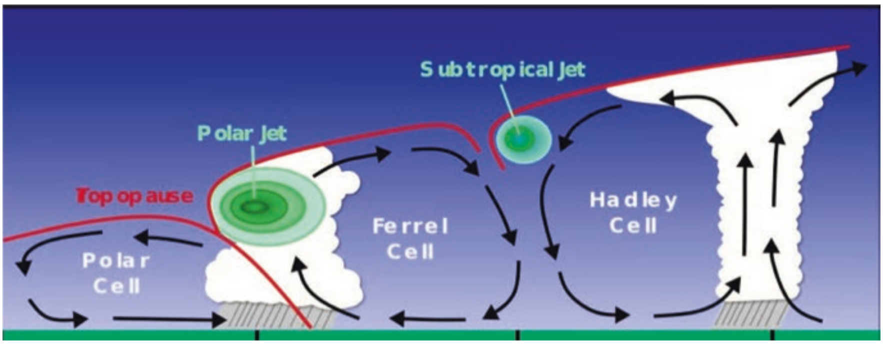


Fig. 15.1 Basic model of atmospheric circulation. (1) Hadley cell; (2) Ferell cell; (3) Polar cell. Credit: Sleske [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)]

basic properties of global pressure and wind patterns (Wells 2012; Schneider and Sobel 2007). The basic assumptions of the model are:

- Uniformed composition of the Earth's surface
- Unequal income of solar radiation, which creates a temperature gradient between the equator and the poles

When these conditions occur on the Earth that rotates, three cells that have characteristic circulation between the equator and the poles are defined:

1. *Hadley cell* between the equator and 30° latitude
2. *Ferrell cell* medium latitudinal between (30° and 60°) latitude
3. *Polar cell* between (60° and 90°) latitude

15.3.1 Circulation in Hadley Cell

The Earth is surrounded by the presence of several widespread prevailing wind belts which are separated by narrow regions on land or regions of air rising. The direction and the location of these belts of wind are determined by the solar radiation and the rotation of the Earth. A more detailed insight in the general atmospheric circulation patterns and the distinct cells is illustrated on Fig. 15.2.

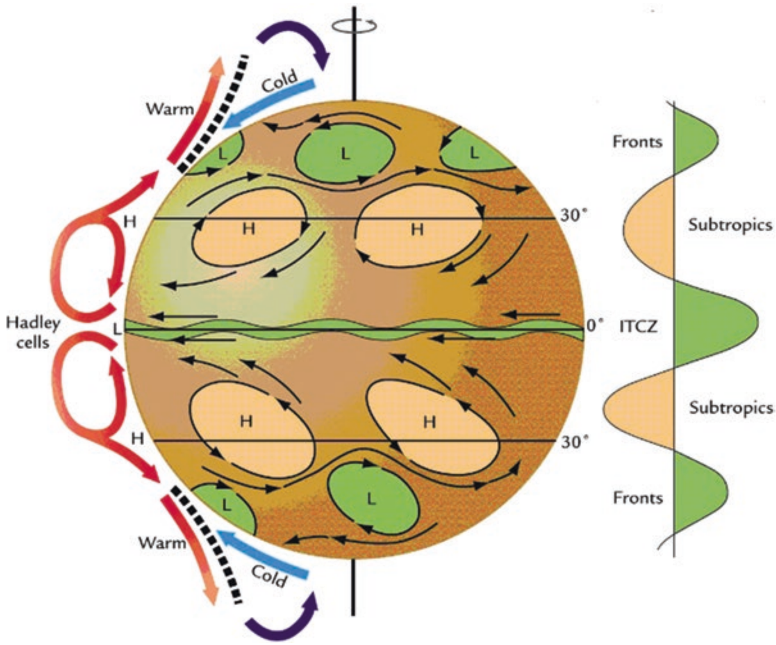


Fig. 15.2 Basic circulation patterns

1. *Hadley cell* – a pattern of atmospheric circulation in which warm air rises near the equator, cools as it travels poleward at high altitude, sinks as cold air, and warms as it travels equatorward also. This is a form of convective cell which dominates the subtropical and tropical climates.
2. *Ferell cell* – average atmospheric circulation cell, whose name was given by Ferell in 1856. In this cell, the air flows towards the poles and the east near the Earth's surface and the equator and westward at higher levels in the atmosphere.
3. *Polar cell* air rises, diverges, and travels to the poles. When there is more than half, air sinks, forming polar centres with high pressure. When it reaches the Earth's surface, air diverges outside the polar zones of high pressure. Surface winds in the polar cell are Eastern polar winds.

At the equator or near it, where the average solar radiation increases, air warms at the Earth's surface and rises. This creates a zone of low air pressure placed on the equator, known as Intertropical Convergence Zone (ITCZ). When this subtropical air reaches the equator, it rises into the upper atmosphere because of convergence and convection. It achieves the maximum vertical height of about 14 km (so-called top of the troposphere), and then it begins to flow horizontally towards the North and the South Pole. This rising air comprises one segment of the circulation model called *Hadley cell*. Hadley cell returns the air to the Earth's surface, close to 30 degrees north and south, respectively. Part with descendent air motion (air sinking) creates Hadley cell zone of high air pressure at these latitudes known as subtropical high. From this zone, the surface air travels in two directions. Winds are generated between equatorial subtropical high- and low-pressure zone, the movement of air from high-pressure to low-pressure area on the surface. The Coriolis force, in combination with an area of high pressure, causes winds to blow towards the equator, forcing them to change their direction from east to west. These prevailing winds are called trade winds or tropical eastern winds (easterlies). Another part of the air from the Earth's surface moves towards the subtropical zone of high pressure. This air is also deviated by Coriolis force, creating western winds.

15.3.2 Intertropical Convergence Zone

The equatorial region of the Earth experiences a total flow of energy throughout the year. More intense heating occurs at lower latitudes, due to the high angle of the Sun and the equal length of day throughout the year. In such case the incoming solar radiation is more direct and concentrated over a smaller surface area, causing warmer temperatures. Heat received by the Earth's surface is transmitted into the air by radiation and by the transfer of latent heat. Condensed water vapour also supplies heat to the surrounding air. Warmer air is less dense and thus easily rises in the wet tropical atmosphere. Convective rise of air that creates a wider area of low pressure surrounding the equator is known as equatorial valley. The equatorial region is called *Intertropical Convergence Zone* (Fig. 15.3). When air is lifting above the

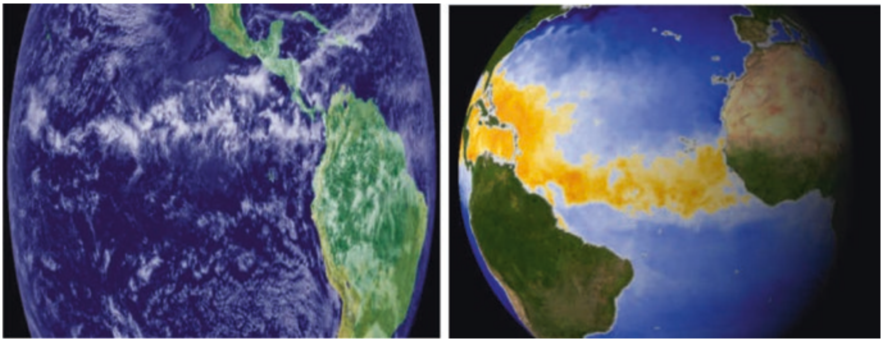


Fig. 15.3 Intertropical Convergence Zone (ITCZ)

Earth's surface, it diverges near the top of the troposphere, and it moves towards equality. Flowing towards the poles, the air begins to converge on (25° and 35°) north and south latitude.

The Intertropical Convergence Zone, or ITCZ, is the region that circles the Earth, near the equator, where the trade winds of the Northern and Southern Hemispheres converge. The intense Sun and warm water of the equator heats the air in the ITCZ, raising its humidity and making it buoyant. The convergence of the trade winds, causes the buoyant air to rise. As the air rises, it expands and cools, releasing the accumulated moisture producing thunderstorms. Subtropical high-pressure belt has a significant impact on the climate of the Earth and represents the driving force in the general atmospheric circulation, much as the major ocean currents. In this zone the downward movement of an air parcel (subsidence) exists as the air cools and becomes more denser forming area of high pressure near the Earth's surface. When the warm compressive air descends, it causes a drop in relative humidity. Our main deserts such as Sahara coincide with the presence of subtropical belts with high pressure. As the air descends towards the surface, it diverges outside the centre of subtropical high-pressure belt. Subtropical high winds in the equator are basically weak and variable. In the Northern Hemisphere, the air is removed from the right path when you move to the outside. Pressure gradient from subtropical high-pressure belt to the equator, between the top 30° N and low over the equator, creates north-eastern trade winds. In the Southern Hemisphere, the air is turned to the left as it diverges from the subtropical high creating the *southeast trades.* Air convergence at the equator occurs between northeast trade winds in the Northern Hemisphere and southeast trade winds in the Southern Hemisphere, creating the Intertropical Convergence Zone. In terms of convergence, trade winds form a zone of calm and weak winds with not dominant (or prevailing constant) direction of the wind called doldrums between 5° north and south latitude. In the subtropical belt of the high-pressure side of the pole, the air is advancing towards the pole, but the flow is removed, to create a form of west wind between (30° and 60°) in the Northern Hemisphere. This zone is known as wind belt with western winds. A similar zone in western winds can be found between 30° and 60° south in the Southern Hemisphere.

The loss of energy at the poles creates very cold air that subsides towards the surface. This creates a dome of high pressure called the *polar high*. Air moving towards the equator is turned in an easterly direction creating the *polar easterlies*. The polar easterlies collide with the westerly wind belt at about 60 N and S creating a broad belt of low pressure called the *subpolar low*. This is a zone of storms and migrating high- and low-pressure systems, the topic of the "Weather Systems" module. Global wind belts and pressure are extremely important elements of the climate system on Earth. They largely determine the geographical pattern of rainfall. Low-pressure systems promote wet conditions, while high pressure tends to prevent precipitation.

15.3.3 Zonal Pressure Patterns

In reality there is only one valid model of the zonal pressure along the subpolar low-pressure zone in the Southern Hemisphere. In other latitudes, especially in the Northern Hemisphere, where there is a higher representation of land than ocean, the zonal model of pressure is replaced by "semi-permanent" cells with high and low pressure. Pressure and wind patterns in January illustrate the strong high-pressure centre, which is known as Siberian anticyclone, positioned above the frozen landscape in northeast Asia. Also evident in January, (absent in July), two intensive "semi-persistent" centres of low pressure and Aleutian's Icelandic depressions extend over the northern Pacific and over North Atlantic, respectively. In the summer, the pressure patterns over the Northern Hemisphere are changing dramatically, and high temperatures occur over the continent, creating depressions that replace winter centres with high pressure. During the summer season, subtropical high-pressure centre, which is in the North Atlantic (called Azorean anticyclone in winter), is positioned near the Bermuda Islands and is known as the Bermuda anticyclone.

15.3.4 Upper Tropospheric Wind and Pressure Patterns

In the previous chapter, we considered the global circulation of air near the Earth's surface. In the high layers of the troposphere, circulation is very different. Between 15 and 20 deg. latitude north and south, located east, elevated winds appear which are considered as a continuation of trade winds. For most of the upper troposphere, moving towards pole, there is an average westward flow, known as the upper western winds. Above the friction layer of the Earth, upper level winds blow with higher speed. Among them, very strong winds are embedded, which surround the entire globe, and they are called jet streams. The strong pressure gradient occurs as the large temperature gradient. Latitudinal temperature gradient of the Earth's surface, and consequently latitudinal gradient of pressure, reaches a maximum in high latitudes. This coincides with the polar front, which marks the boundary between polar and tropical air type. Here lies the region of the polar jet stream front, high-speed corridor of wind that is responsible for the creation and movement of large systems with pressure (baric systems) across the middle latitudes. Additionally, subtropical jet stream is formed at approximately 30° latitude, as a result of the convergence of air at altitude, the same width as the one discussed earlier. At the poles, air is very cold and subsides (sinks), to form a field with high air pressure at the surface. Entry creates a huge area of low pressure at an altitude known as circumpolar vortex. A jet stream of the polar front lies close to the equatorial border of the vortex. Polar vortex expands and shrinks during the year as a result of energy flow shifting to the surface. Equatorial edge, along jet stream, takes a wave shape with cold air, which penetrates high latitudes, and warm air blows to pole. In these circumstances, there is a transfer of cold air to the south and a transfer of warm air northward. In such situation, polar vortex experiences a meridian flow. Sometimes the edge is flatter, blowing more evidently from west to east in the direction called zonal flow. In this scenario there is little latitudinal exchange of air and energy. This brings relatively mild maritime polar air masses over the continents, the lack of expression temperature difference.

15.4 The Wind Patterns

Depending on the horizontal scale of motion, certain scales of wind patterns are defined. The global wind pattern is also known as the "general circulation" (Fig. 15.4). The surface winds of each hemisphere are divided into three wind belts:

Polar easterlies: From 60 to 90 degrees latitude
Prevailing westerlies: From 30 to 60 degrees latitude (westerlies)
Tropical easterlies: From 0 to 30 degrees latitude (trade winds)

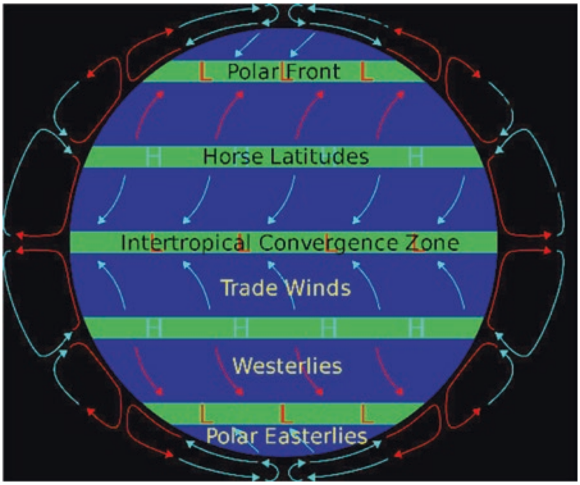


Fig. 15.4 Permanent winds of the general atmospheric circulation. Credit: DWindrim (2004-02-08)derivative work: Burschik (2007-05-31) [CC BY-SA 3.0 (http://creativecommons.org/licenses/by-sa/3.0/)]

The circulation which has lower horizontal scale resolution than macro-scale circulation is called synoptic-scale circulation or scales of weather maps. Winds associated with the mesoscale processes and phenomena, such as stormy winds, gusts land, and sea winds, affect smaller areas and experience the intense vertical movements. The smallest (spatial and temporal dimensions) air motion represents a circulation in a microscale. Examples of these local, often chaotic winds are tornadic winds, turbulence vortices, and sandstorms.

15.4.1 Eastern and Western Winds

The circulation between (30° and 60°) latitude (north and south) results in prevailing western winds. The air that moves from the poles to the equatorial polar regions creates eastern winds on both hemispheres.

The area where the cold polar east winds collide with the warm west flow in high latitudes is known as the polar front, and it is an important meteorological region. If the Earth's surface is uniformed, two belts of high pressure and two belts of low pressure will exist, parallel to the equator. Starting from the equator, these four belts are:

1. Equatorial belt of low pressure, known as Intertropical Convergence Zone (ITCZ), because it is a region of trade winds convergence
2. Subtropical zone with high pressure, about (20° to 35°) width on each side of the equator
3. Subpolar low-pressure zone, which is about 50° to 60° latitude
4. Polar high-pressure zone, near the poles

15.5 Global Distribution of Pressure, Rainfall, and Climate

Global atmospheric pressure systems play a direct role in the geographical distribution of rainfall. This influence is obviously the relationship between atmospheric models and distribution of rainfall and climate. In general, there is a horizontal pattern of climates that extends from the equator poleward; at 30° N is associated with the location of Intertropical Convergence Zone (ITCZ), semi-permanent subtropical high-pressure zone (STH), and cyclones in high latitudes. At the equator prevails a climate with the greatest amount of rainfall on the Earth, tropical rainy forest (AF), and tropical monsoon (Am). According to recorded data, the annual amounts of rainfall often exceed (2540 mm or l/m^2) in the year. Heavy rains occur as a result of warm moist air masses which converge in the low-pressure belt, extending over the equator and rising through convection. Monsoon climate in Africa and Asia is evident during dry periods of weak expressed sunny season. The dry period is due to the presence of air down from the subtropical high-pressure belt (STH), while the wet season is due to the presence of the ITCZ and moisture to carry trade winds along the coast. From the monsoon climate and the rainy forest environment in the pole, tropical wet/dry climate (Aw) results, often known as “Savannas”, climate. Like the monsoon climate, rainfall is seasonal, with much smaller amounts of annual rainfall, which receives tropical wet/dry climate. During low Sun season (in winter), tropical wet/dry climate is influenced by subtropical high-pressure belt (STH), causing extremely dry conditions. Savannas climate is considered a climate that is transitional between rainy and dry tropical climate. Tropical steppe climate (BS) is dominated by STH in most of the year, through which it receives a high Sun season, when the ITCZ rain moves into the region. From the Steppes to the pole, tropical desert climate (BW) exists. The tropical desert climate is the driest on Earth, so in some places there is no measurable precipitation on an annual basis. The extreme aridity is due to the impact of subtropical zone of high air pressure throughout the year. At about 30° north latitude, there is another area where tropical steppe climate occurs. While climatic conditions are the same, the rule of seasonal baric systems is reversed. During low Sun STH season it moves into the region to create dry conditions. During the winter cyclones that produce high latitude occasional short rains occur. Finally, the Mediterranean manifests dry subtropical climate summer (Cs), or sometimes known as “Mediterranean climate”. Similar to steppe climate of equatorial border of the Mediterranean climate, there is a summer/wet winter rainfall regime. However, the region receives more rain annually than neighbouring steppe climate.

15.6 Ocean Circulation

Ocean circulation affects the energy balance of Earth, the general atmospheric circulation and interactive processes that take place in the atmosphere, land, and water surface (Wells 2012). About half of the transport of heat around the Earth’s planet is going through the oceans, so the oceans have an extremely important role in controlling the Earth’s climate system. The energy of the Sun falls evenly over the entire surface of the Earth. Most of the influx of solar radiation that travels the Earth occurs at the equator. This leads to large temperature gradients between the equator and poles. The flow of air and oceans is controlled by the temperature differences, and it occurs as a result of the transfer of heat from the equator to the poles.

Figure 15.5 shows Earth’s sea surface temperature obtained from 2 weeks of infrared observations by the Advanced Very High-Resolution Radiometer (AVHRR), an instrument on board NOAA-7 during July 1984. To understand the reason for these temperature anomalies, we need to understand ocean currents. A part of the large-scale circulation of water through the oceans is called the thermohaline circulation. It is driven by global density gradients formed by surface heat and ocean water fluxes. If the ocean circulation changes through global warming, major changes in climate are likely to happen. The thermohaline circulation is also known

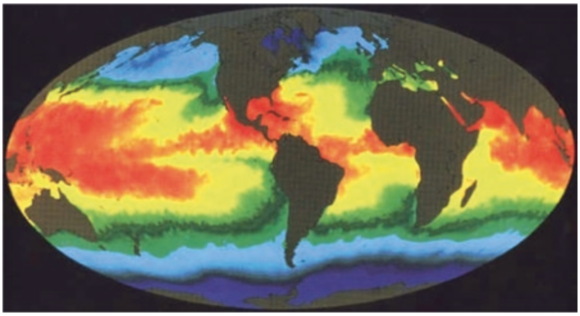


Fig. 15.5 Global sea surface temperature (SST)

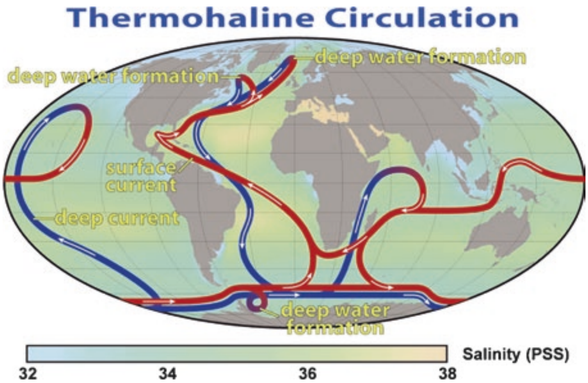


Fig. 15.6 Basic model of global ocean circulation. Credit: Robert Simmon, NASA

as "meridional overturning circulation" (see Fig. 15.6) which is responsible for the large-scale exchange of water masses in the ocean, including providing oxygen to the deep ocean. There is a clear correlation between the pattern of ocean currents and atmospheric circulation. Similar to the atmospheric circulation, the ocean circulation is significant in the latitudinal transport and redistribution of energy. Ocean behaves as a heating system which is like the atmosphere, but the ocean is considerably less efficient. This is because the sources and sinks of heat in the oceans are near the surface, even at approximately the same air pressure. The largest movement of the ocean, especially at the higher layers, is a result of wind friction near the surface. As the air moves faster through the ocean surface, water resistance reduces the air flow in the layers close to the surface. As a result of these interactive processes that take place between the atmosphere and oceans, friction occurs at the interface, which comes in the form of turbulent movements in the atmosphere. The pattern of thermohaline circulation also known as "meridional overturning circulation" resistance (friction) is most dominant in winter when the winds are strongest, and it shows a different annual cycle. The wind friction is a very important part of the atmosphere-ocean interaction and is a key link in the whole climate system. It is also important to recognize that currents associated with wind are limited to the layer of water that has a uniform temperature distribution and is known as the mixing layer with an average thickness of about 70 cm.

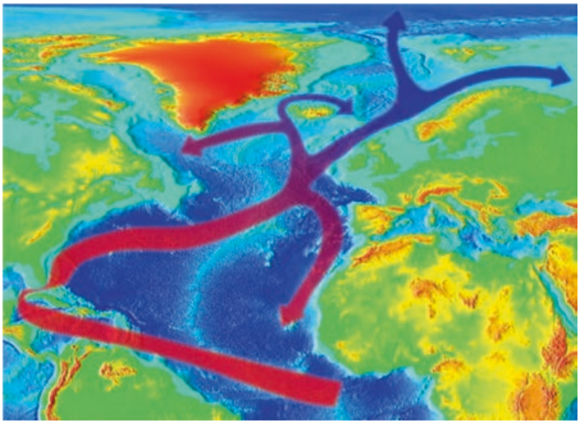


Fig. 15.6 Gulf Stream map. Credit: RedAndr [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)]

15.6.1 *Major Ocean Currents*

The main ocean currents are generated by wind. Some ocean currents result from the changes in density and salinity of the water. Subtropical high-pressure cells are responsible for many of the larger Earth ocean currents. Ocean influence on climate is mostly manifested in the heating and cooling in the coastal areas. Common to them is that the great ocean currents are classified as warm or cold, and their global distributions are shown on Fig. 15.6.

15.6.1.1 Warm Currents

One of the most significant ocean currents which occur in the northern Atlantic is the Gulf Stream (Fig. 15.6). It is a large, warm current that flows northeast between Cuba and southern Florida, in parallel with the coast of Northeast America, almost to Newfoundland, then continues as Northeast current.

The northeast ocean stream then moves east, passing the United Kingdom and Norway on the eastern end of Scandinavia and comes in Arctic Barents Sea north of Russia. The Gulf Stream has a profound effect on most of the European continent, because there are a few orographic barriers at these latitudes, which retain the western winds as bearers of sea air to the east. In the Norwegian Sea in winter, Gulf Stream causes ground temperature that can be more than (26°C) above the normal value for this latitude. Experiences show that winters in most of the British island, which lies north of the breadth of the Canadian border, are relatively mild. Similar impact on the Gulf Stream is manifested in George, which is (15°C) to (20°C) latitude farther south. The North Pacific current, which is an extension of the Kuroshio current, has no strong effect on climate on the east side of the Pacific, compared to the Gulf Stream on the east side of the Atlantic, because its impact must prevent spreading towards the pole. High mountains that are parallel to the west coast of

North America do not allow penetration of the great Pacific sea air across the continent. Also, the northern Pacific is separated from the Arctic in winter by the narrow Bering Marine freeze, which blocks any movement of surface water from the North Pacific in the Arctic, middle and late winter. Warm currents are weekly developed in the Southern Hemisphere. Indian Ocean flows largely under the influence of seasonal monsoons. Weak, diffuse warm currents moving southwest along the east coast of Australia.

15.6.1.2 Cold Currents

Ocean currents with the most dominant climate impact on the Southern Hemisphere large cold currents. These are major cold currents in the oceans. Peruvian stream has the highest volume and most important climate, currents north along the west coast of South America around (50° S), almost to the equator. This stream has a significant effect on the climate along most of the west coast of South America, although its influences are limited to a narrow coastal section of the high Andes. Cold water in the stream due to springing, in response to movement of surface water by winds near the coast. Cold ocean water and atmospheric subsidence of subtropical high-pressure air, create a strong temperature inversion. When these conditions occur, there is quite a poor rainfall. As a result of these processes, the coast is characterized by the appearance of cold, moist surface air, with frequent occurrence of fog and low coverage of stratus clouds. Bengal stream flows north along the southern coast of Africa, the eastern South Atlantic. Cold currents in the Northern Hemisphere are not strong. Canary's stream flows southwest, along the banks of the Iberian Peninsula and the northwest coast of Africa, passing the Canary Islands. California stream flows southwest along the west coast of North America. During the summer, the water away from the coast north of San Francisco is often coolest along the entire North American west coast. The local wave increases are more significant than the horizontal movement of water from Alaska to the south. Comparing the temperature of the water surface in late August illustrates the nature of the cold stream in relation to the warm current (Gulf Stream) along the east coast.

15.7 Ocean Waves

The waves that occur on the surface of the sea and ocean are very significant for ocean circulation. They transport a horizontal momentum to the Ekman layer, accelerating vertical mixing in the upper layers of water. Waves play a prominent role in the exchange of gas between the atmosphere and the ocean. They mediate carbon dioxide into the oceans and oxygen into the atmosphere. Winds above the ocean surface not only create ocean currents but also waves. When the wind blows above the smooth surface of the water, friction occurs between air and water that seeks to stretch a thin layer of water like that of a soapy water can. Vortex formation and wrinkling, browning, water surfaces can occur at some point. The slightly uneven surface of the water (the minimum wave propagation velocity is 0.23 m/s) allows the wind to more easily form the waves that accompany it. The wind becomes turbulent just above the surface of the water. By acting on increasingly steady surface of the ocean, the wind raises larger and larger waves. If the speed of the wind is greater than the speed of wave propagation, it transfers the energy to the surface of the ocean. From the point of view of the action of forces, such waves arise because of the force of Earth's gravity, which acts in the sense of restoring hydrostatic equilibrium in the ocean. Disturbance on the surface of the water in terms of height (depth), a pressure gradient force occurs that causes the water to accelerate in a horizontal direction from a place of greater water depth to less depth. Therefore, such waves are called gravitational waves. The point of the highest wave height is the wave front, and the lowest is the valley height. Wave height is the total vertical change in height between the bank and the valley. The distance between two consecutive waves (or valleys) of a wave represents the wavelength of the wave. When passing the bank of the wave, the particles the waters describe the trochoid (a curved line describing the trajectory of a point on a rolling circle). This means that the wave not only produces oscillations of the particles along circular paths on the surface, but there is little movement of the particles in the direction of the wave propagation, especially at large ones. These almost circular paths of water depth sections become more ellipsoidal and smaller in size until they completely disappear at approximately the depth corresponding to the adjacent wave valley. For deep ocean waters, the speed of propagation of gravitational waves depends on the wavelength. The greater the wavelength, the greater the speed of wave propagation. So, for example, a tsunami with a wavelength of 100 km and above can have a speed of almost 900 km / h. For deep oceans, the wavelengths of surface waves can be no more than twice their mean depth. For shallow water, the speed of gravity wave propagation depends on the depth of the water. The greatest waves can be produced by strong and long winds, which all time is blowing in one direction above a certain area. Strong winds are more turbulent and form waves more easily. Extremely high waves are observed in the Indian Ocean in the band between 40 ° and 60 ° in the Southern Hemisphere (Fig. 15.7). The mean wave height was about 7 m, while some waves had twice the height. South of New Zealand the wave height was about 5 m. Area about equatorial coils are characterized by weak winds and therefore low waves, while islands in the tropics greatly limit the wind to blow in a constant direction. However, huge waves below the typhoon or hurricane can also be raised in this area. So far, there has been talk of waves on the surface of the ocean that are gravity waves character. Otherwise, we have progressive and standing waves in the oceans. Progressive include surface, internal, and tsunami. Sound waves propagate through water about 4.5 times faster than through the atmosphere. Depending on the temperature and salinity of the water, their velocity is approximately 1500 m/s. They extend in all directions. If a sound wave is emitted on the surface of the water, it can reach the bottom of a depth of 4 km in just a few seconds. The lowest waves occur where wind speeds are lowest, around the equator, indicated by the pink colour on this map. However, in these places, the sea water warms up, causing the birth of

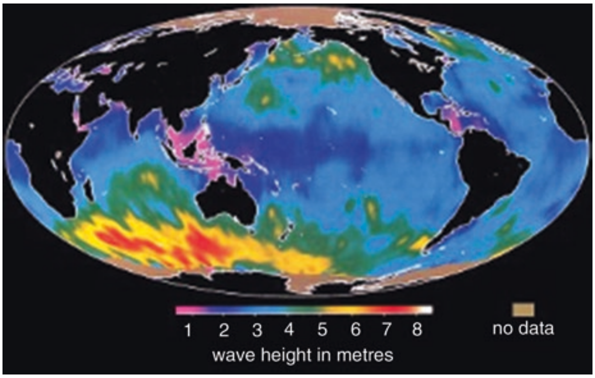


Fig. 15.7 Global distribution of ocean wave altitude

tropical cyclones, typhoons, or hurricanes, which may send large waves in all directions, particularly in the direction they are travelling.

15.8 Large-Scale Circulation Modes

It has been found that Sir Gilbert Walker in the 1920s in his work described three major large-scale oscillations within the general atmospheric circulation:

- The Southern Oscillation in sea level pressure across the equatorial Pacific Ocean
- The North Atlantic Oscillation between the Azores high-pressure cell and the Icelandic low
- The North Pacific Oscillation between the North Pacific High and the Aleutian Low

15.8.1 North Atlantic (NAO) and Artic Oscillation (AO)

The term "North Atlantic Oscillation" refers to variations in the large-scale surface pressure gradient in the North Atlantic region, i.e. (low pressure around Iceland/high pressure around the Azores). Another related mechanism focuses on "storm tracks". These are the relatively narrow paths over land and sea through which storm weather systems tend to be pushed by prevailing winds. In the Northern Hemisphere, the position of storm tracks is affected by a natural fluctuation known as NAO. This natural weather phenomenon consists of two pressure centres in the North Atlantic: one is an area of low pressure typically located near Iceland and the other an area of high pressure over the Azores (an island chain located in the eastern Atlantic Ocean).

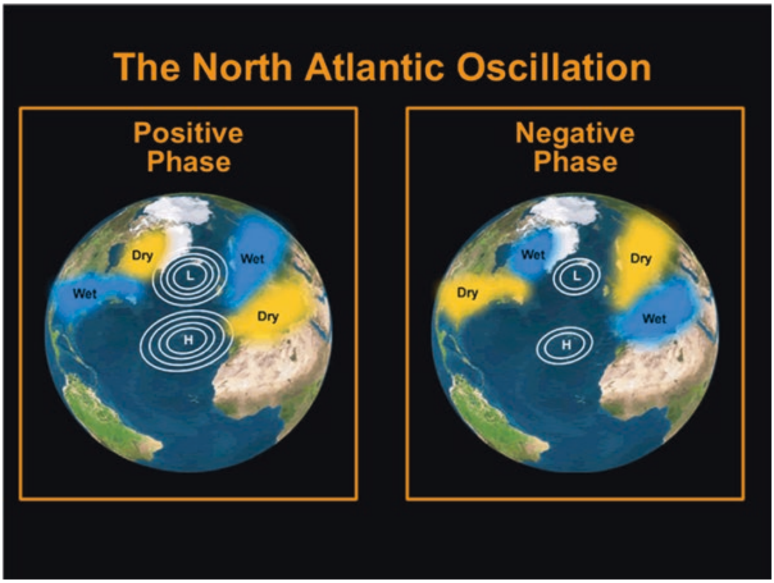


Fig. 15.8 **Negative** and positive phases of NAO. Courtesy: UCAR

Figure 15.8 shows a different NAO phases. When the pressure difference is low, NAO is negative. When both the sub-polar low and the subtropical high are stronger than average positive phase of NAO is dominant. The NAO is also associated with changes in temperature and rainfall in Europe and North America.

When the pressure difference is large, the NAO is positive, and the westerly winds are strong, and storms tend to be stronger, more frequent, and travel across north Western Europe. Fluctuations in the strength of these features significantly alter the alignment of the jet stream (discussion follows) and ultimately affect temperature and precipitation distributions in this area. AO is a climate index of the state of the atmospheric circulation over the Arctic (Fig. 15.9). It consists of a positive phase, featuring below average geopotential heights, which are also referred to as negative geopotential height anomalies and a negative phase in which the opposite is true. In the negative AO phase, the polar low-pressure system (also known as the polar vortex) over the Arctic is weaker (see Fig. 15.10), which results in weaker upper level winds (the westerlies). The result of the weaker westerlies is that cold, Arctic air can push farther south into the US continent and Europe, while the storm tracks also remains farther south. It is also important to note that the AO and NAO are two separate indices that are ultimately describing the same phenomenon of varying pressure gradients in the northern latitudes and the resultant effects on temperature and storm tracks across the continent. There is debate over whether one or the other is more fundamentally representative of the atmosphere's dynamics.

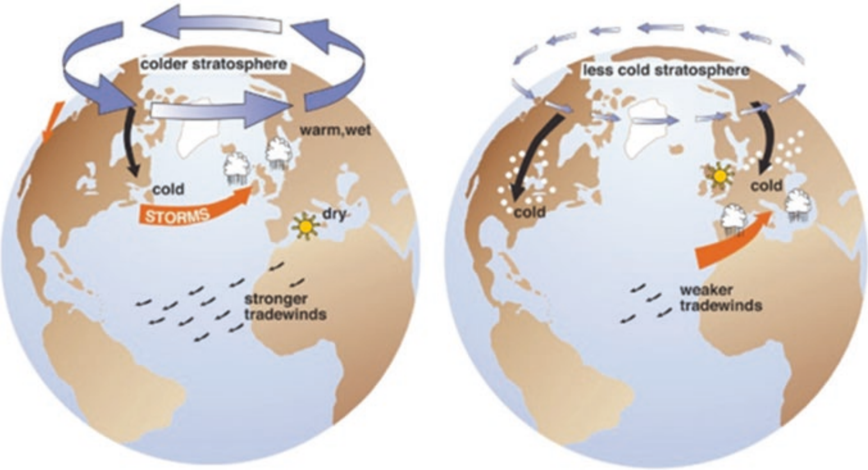
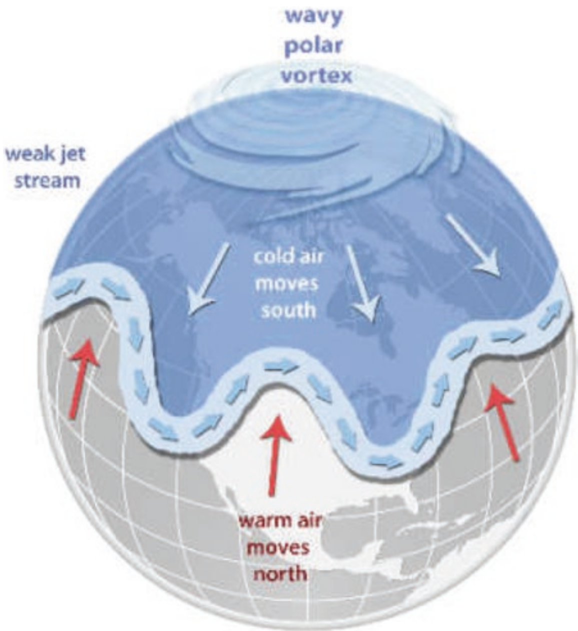


Fig. 15.9 Artic Oscillation

Fig. 15.10 Polar vortex



15.8.2 El Niño-Southern Oscillation

The phenomenon “El Niño”-Southern Oscillation is a major climatic disturbance, generated in the tropics of the Pacific, and it occurs every 3 to 7 years. This phenomenon has a strong influence on the continents around the tropical Pacific, and there is some climatic influence on half of the planet Earth. The development of El Niño-Southern Oscillation is characterized by raising the temperature a few degrees, the ocean surface (3 to 6 °C), and the coasts of Peru and Ecuador to the centre of equatorial Pacific Ocean. Because of this warming, long-term perturbations of weather systems over the surrounding land, expressed heavy precipitation in normally arid areas, drought in the normally moist areas occur. El Niño-Southern Oscillation is also observed as a hot phase of irregular climatic oscillation, known as ENSO (El Niño-Southern Oscillation/Southern Oscillation), which is caused by the unstable interaction of the ocean and atmosphere (Fig. 15.11).

15.8.3 La Nina

Conversely, “La-Nina” is the cold phase of ENSO recurring climate pattern. During the “La Nina” episode trade winds strengthen, warm water and rainstorms are shifted far away towards western equatorial Pacific Ocean, resulting in cooling of surface waters and below averaged sea surface temperature in the equatorial Pacific Ocean. (Fig. 15.12). La Nina case may follow El Niño-Southern Oscillation, but not always.

When surface sea surface temperatures in the central and eastern Pacific Ocean are lower than average, appears a phenomenon called “La Nina”. It represents typical winter blows colder than constant air over the north-western Pacific and over northern high plains, simultaneously heating the rest of the American continent.

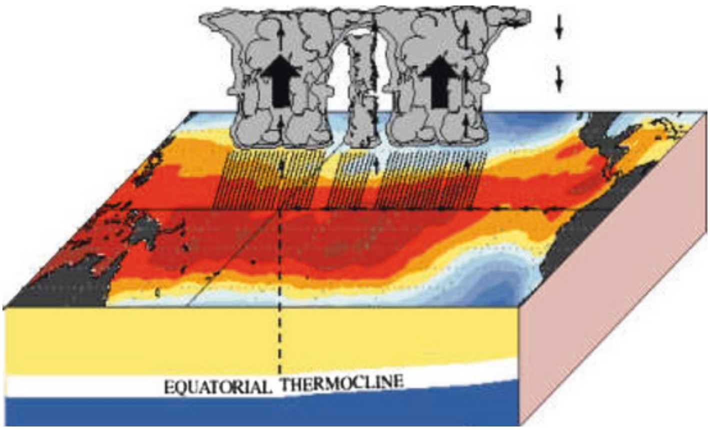


Fig. 15.11 El Niño (warm phase)

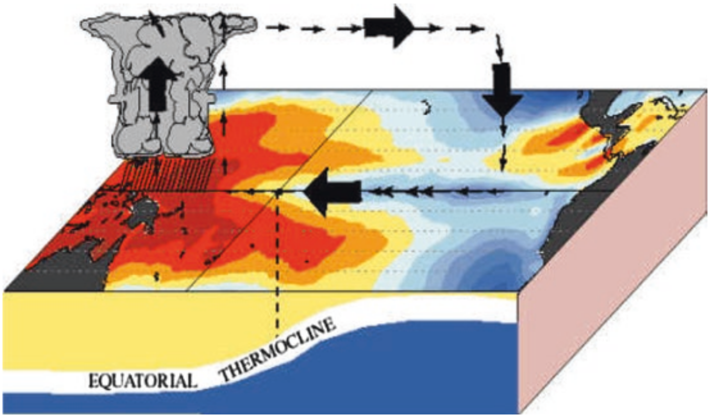


Fig. 15.12 La Nina (cold phase)

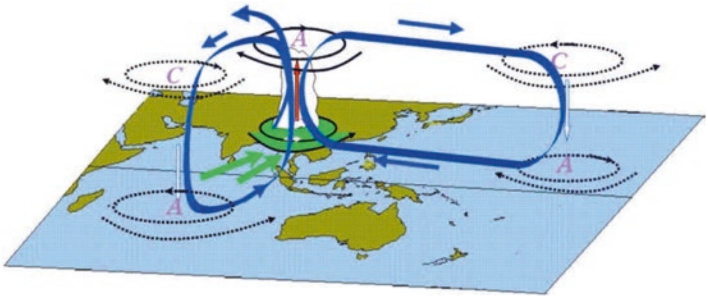


Fig. 15.13 Summer Indian monsoon circulation. Credit: Chang (2004)

15.9 Winds at the Regional Scale: Monsoon

The largest seasonal change in global circulation on Earth is the development of monsoons, wind systems that experience highly seasonal deviation in direction. The term "monsoon" means wind changing direction with season. Monsoon conditions are found in different regions around the world. European atmospheric circulation induced by temperature differences between the continent and the North Atlantic Ocean causes thermodynamic and climatic conditions that initiate a European monsoon. As it is illustrated in Fig. 15.13, most notable is the characteristic wind pattern over Asia, called Indian monsoon (Clift and Plumb 2008; Chang 2004; Chang et al. 2011). During summer, the continent of Asia is more heated than the surrounding ocean, as a result of the asymmetric differential heating of land and sea. Heated surface forms a large area of low pressure in the northern part of Central Asia and smaller area than India. This creates a coastal wind that carries the mainland marine air rich with moisture from the Pacific and Indian Ocean. As air flows through the land, warm surface converges in areas with low pressure. It causes air rising and formation

Fig. 15.14 Weather conditions in monsoon season

of strong updrafts along the major mountain systems such as the Himalayas, where severe thunderstorms are initiated with a heavy rainfall during the wet monsoon season. In winter, the air flow is opposite. Continent rapidly cools forming a large area of high pressure over north central Asia, known as the Siberian Anticyclone, and less area than India. Drier and colder air blowing out of the continent from the coast is creating monsoon dry season.

Monsoon conditions also manifest in eastern central Africa. This region experiences seasonal rains, caused by the variation influence of global atmospheric pressure systems. During the high Sun season, the ITCZ moves inside carrying a warm and moist unstable air-conditioned heavy rain (Fig. 15.14). During low Sun season, the ITCZ moves to the outside, and the impact of descendent movements of subtropical high-pressure air (anticyclone) dominates the region. During this period, subtropical anticyclone prevents the development of precipitation.

Deserts southwest of the North American continent experience pseudo-monsoon during the summer. This monsoon (often called Arizona monsoon and south-western monsoon) is a relatively small seasonal shift of wind, which creates a dry spring, followed by comparatively rainy summer, affecting large areas of the southwest of the continent and north-western Mexico. Intense daytime heating of the surface depression (low pressure), creates a favorable conditions for initiation of thunderstorms and rainfall over the desert areas, typically occurring between June and September. During winter the desert area is cooling, low pressure centre disappears, and dry conditions prevail. It is important to note that there is a European monsoon in eastern and Western Europe as a result of the temperature difference between the northern Atlantic Ocean and the continent. Therefore, it is rainy early summer in Eastern Europe and rainy early winter in Western Europe. Dry periods occur in winter in Eastern and in spring in Western Europe. Due to the presence of monsoons, large floods are frequent in some parts of central Europe (Radinovic and Ćurić 2012).

15.10 Jet Streams

Jet currents are relatively narrow belts of strong winds in the upper levels of the atmosphere (Shapiro and Keyser 1990). Winds blow from west to east in jet streams but flow often moved to the north and the south. Jet streams follow the boundaries between warm and cold air. Because these warm and cold air boundaries are mostly highlighted in the winter, jet streams are strongest in the Northern and the Southern Hemisphere during the winter. The temperature difference between the equator and the poles is causing western winds located in high latitudes. The narrow belt of high-speed winds, trapped inside the western flow of height, covering thousands of kilometres, is called jet streams. Jet streams in high latitudes arise as a result of large temperature differences on the Earth's surface. In the middle latitudes between 30° and 70°, smooth polar strength accompanied by a polar front (Fig. 15.15) appears. Because the slopes of cyclone systems are managed by the flow of the height, position of the polar jet stream has a strong influence on the weather, especially temperatures, the surface. Because winds are the basic driving force of ocean currents, it reveals a link between ocean circulation and atmospheric general circulation. In general, as response to the circulation associated with subtropical high-pressure centres, ocean currents form a spiral clockwise in the Northern Hemisphere and counterclockwise in the Southern Hemisphere.

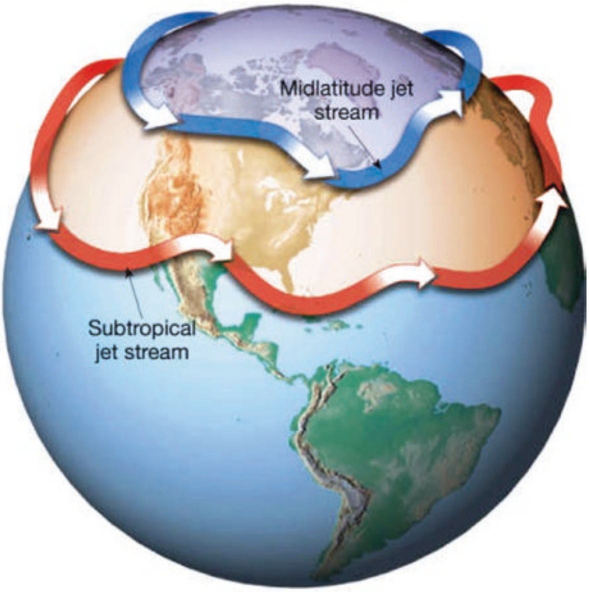


Fig. 15.15 Polar and subtropical jet stream. Credit: https://en.wikipedia.org/wiki/Lyndon_State_College_Meteorology

15.10.1 A Waiver Jet Stream

Fluctuations in the strength of the NAO and AO patterns significantly alter the alignment of the jet stream and ultimately affect temperature and precipitation distributions in this area. The theory goes that these changes contribute to an increase in unusual and extreme weather across the North America, Europe, and Asia. The different theories of how Arctic amplification could be affecting the mid-latitudes fall broadly into three main categories: those that focus on a "wavier" jet stream, the stratospheric polar vortex (sudden stratospheric warming), and "wave resonance" relate to the jet stream in the troposphere, which only occurs during summer and can trigger extreme weather (e.g. heatwave, severe storms, and floods). According to this theory, the troposphere is the most significant – lowest layer of the atmosphere, where most of the day-to-day weather occurs.

In particular, it concerns the polar "jet stream" – a band of fast-flowing air high up in the atmosphere. The strength of the jet stream is driven by the difference in temperature between the cold air over the Arctic to the north and the milder air to the south. Since the Arctic is warming more rapidly than the mid-latitudes, the temperature difference is declining, which leads to a weaker jet stream. "A weaker jet stream is more easily deflected from its generally west-to-east trajectory by obstacles in its path, such as mountain ranges, varying sea surface temperature patterns and injections of energy from dying tropical storms". These diversions from its path increase the likelihood of a wavier jet stream pattern, says Francis. When the jet stream takes larger north-south meanders – known as "Rossby waves" – warm air can penetrate north, and cold air can plunge south. Larger waves also mean the systems moving from west to east tend to travel more slowly, "effectively making weather conditions more persistent". "When a particular weather pattern stays in the same place for a long time, it can become an extreme event – think drought, prolonged rainfall leading to flooding, persistent cold spells, and long-lived heat waves".

15.11 Rossby Planetary Waves

Oceanic and atmospheric Rossby waves (Fig. 15.16), also known as planetary waves, naturally occur largely due to the Earth's rotation. These planetary (ocean and atmospheric) waves play a significant role on the planet's weather and climate. Slow-moving oceanic Rossby waves are significantly different from ocean surface waves. Rossby waves are large, disturbing movements of the ocean that extend horizontally across the planet for hundreds of kilometres in a westward direction. Along with rising sea levels, oceanic Rossby waves contribute to high tides and coastal flooding in some regions of the world.

The horizontal speed of Rossby waves is dependent upon the latitude of the wave. In the Pacific, for instance, waves at lower latitudes (closer to the equator) may take months to a year to cross the ocean. The vertical motion of Rossby waves

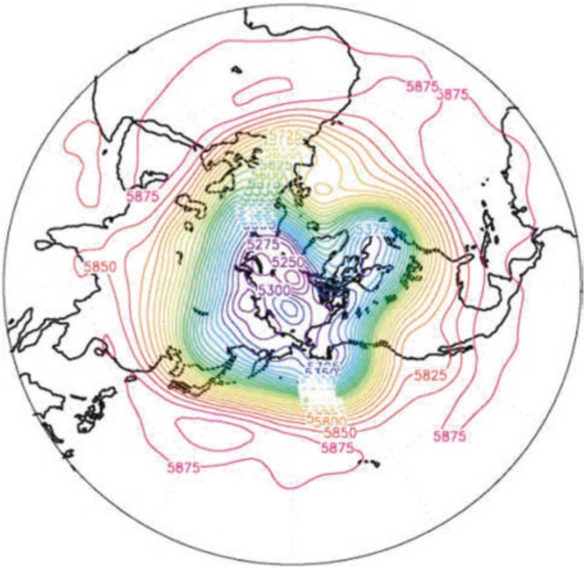


Fig. 15.16 Rossby waves generated by a source with vortex oscillation. Credit: Puncinus [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)]

is small along the ocean's surface and large along the deeper thermocline – the transition area between the ocean's warm upper layer and colder depths. This variation in vertical motion of the water's surface can be quite dramatic: the typical vertical movement of the water's surface is generally ~0.1 m, while the vertical movement of the thermocline for the same wave is approximately 100 m or greater. Due to the small vertical movement along the ocean surface, oceanic Rossby waves are difficult to observe. Scientists typically rely on satellite radar altimetry to detect the massive waves. Atmospheric Rossby waves form primarily as a result of the Earth's geography. Rossby waves help transfer heat from the tropics towards the poles and cold air towards the tropics to return atmosphere to balance. They also help locate the jet stream and mark out the track of surface low-pressure systems. The slow motion of these waves often results in long, persistent weather patterns.

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Chapter 16
Air Masses and Fronts

16.1 Definition of Air Masses

Synoptic map analysis, Swedish meteorologist Tor Bergeron (Bergeron) found that in the air above the vast areas, the values of certain meteorological elements change very little in the horizontal direction. He called these large masses of air masses of air. The horizontal dimensions of the air masses are of continental scale (from 500 to 5000 km), while in the vertical direction, they can cover the entire troposphere. Air masses are characterized by temperature, humidity, temperature stratification, and degree of turbidity. Large, synoptic air masses will obviously be discussed here. It should be noted that meteorology also refers to smaller air masses. Air mass is defined as a widespread body of air with uniform physical properties situated over a particular region of the Earth's surface called air mass source region (e.g. Ahrens 2007; Lupikasza 2000; Hoskins and James 2014; Lackmann 2012). It represents a large volume of air with temperature and moisture regime that tend to be relatively homogeneous in the horizontal direction over wide area. In addition, the vertical temperature and moisture variations are approximately the same over its horizontal extent. Air masses usually undergo specific modifications while in transit away from the source region. The stagnation or long-continued motion of air over a source region permits the vertical temperature and moisture distribution of the air to reach relative equilibrium with the underlying surface. The air mass is separated from the adjacent air mass through the border area, which might be more accurately defined. Where two air masses of different temperatures meet, a boundary forms which is termed a "front". The air mass is covering thousands of square metres, and it extends vertically through the troposphere.

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16.1.1 Air Mass Source Region

The temperature of an air mass will depend largely on its point of origin and its subsequent displacement over the land or sea. This might lead to warming or cooling by the extended contact with a warm or cool surface. Warm air mass is produced by prolonged contact with a warm surface, and conversely a cold air mass is produced by prolonged contact with a cold surface. The heat transfer processes that warm or cool the air take place slowly. It may take a week or more to warm up the air by $10^{\circ}C$ right through the atmosphere, and for these changes to take place, a large mass of air must stagnate over a region. Parts of the Earth's surface where the air can stagnate and gradually gain properties of the underlying surface are called source regions. The main source regions are the high-pressure belts in the subtropics (giving rise to tropical air masses) and around the poles (the source for polar air masses).

16.1.2 Formation Criteria

Air mass source regions are geographic areas where an air mass originates. Two criteria should be satisfied for air masses formation:

- Large and physically uniformed
- Air mass stagnation (presence of light surface winds)

The longer the air mass stays over its source region, the more likely it will acquire the properties of the surface below.

16.2 Air Mass Classification

Air masses are classified into different types primarily by the area in which they originate and depending on their basic temperature and humidity characteristics (Hoskins and James 2014). Also, the classification depends on the distribution of the source region and the nature of the surface of the source region (continent-ocean). The Bergeron classification is the most widely accepted form of air mass classification. Air mass classification involves three letters.

The first letter describes its moisture properties, with **c** used for continental air masses (dry) and **m** for maritime air masses (moist). Its source regions are denote with T for tropical, P for polar, A for Arctic or Antarctic, M for monsoon, and E for equatorial. The generalized map of global air masses is illustrated on (Fig. 16.1).

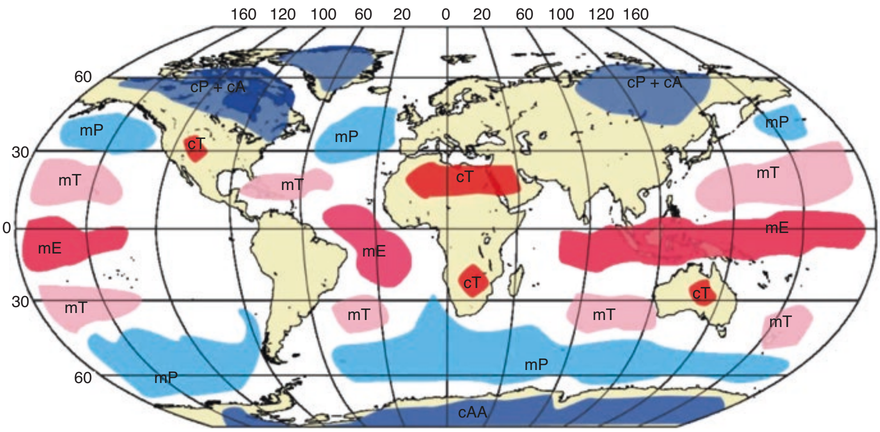


Fig. 16.1 Source regions of common air masses (Bergeron classification). Credit: Public Domain, https://commons.wikimedia.org/w/index.php?curid=12526643

16.2.1 Polar Continental and Artic Continental Air Masses (cP and cA)

These air masses have its origins over the snow fields of Eastern Europe and Russia and are only considered a winter (November to April) phenomena. During the summer with the land mass considerably warmer, these air masses would be classed as a tropical continental. Daily manifestation of weather depends on temperature stability and moisture content of the air masses. Polar continental (cP) and Artic continental (cA) air mass as their classification indicates is cold and dry. Although cP-air masses are not strictly associated with heavy rainfall, those who pass the Great Lakes during the late autumn and winter sometimes bring snow to the effects at the upwind side of coast on the lake.

16.2.2 Maritime Polar Air Masses (mP)

Polar maritime air masses (mP) are formed over the ocean at higher latitudes, and they are wet and cool, to cold. Instability is caused by torrential rains over the sea and permanent rain in exposed western and northern British Isles. During the winter months, when convection is initiated over the sea, hail and thunders are common in these mountainous areas. This air mass has its origins over northern Canada and Greenland and reaches the British Isles on a north-westerly air stream. In summer months, temperatures on the mainland are higher than the temperature of the sea, and intense showers occur over the eastern part of the British Isles. This air mass starts very cold and dry, but during its long passage over the relatively warm waters of the North Atlantic, its temperature rises rapidly, and it becomes unstable to a great depth. This air mass is characterized by frequent showers at any time of the year. In the winter months when instability (convection) is most vigorous over the sea, hail and thunder are common across much of the western and northern side of the British Isles. However, eastern Britain may see fewer showers as here the surface heating is reduced. During the summer, the reverse is true, land temperatures are higher than sea temperatures, and the heaviest showers occur over eastern England.

16.2.3 Tropical Maritime Air Masses (mT)

Tropical maritime (mT) air masses, which cover North America, form over the warm waters of the Atlantic Ocean, Gulf of Mexico, Caribbean Sea, and between the Azores and Bermuda. Tropical maritime air is warm and moist in its lowest layers and, although unstable over its source region, during its passage over cooler waters becomes stable and the air becomes saturated. During the winter, when the cP air dominates the central and eastern parts of the United States, tropical maritime air only occasionally enters this part of the country. However, the summer-mT air masses from the Gulf, Caribbean, and neighbouring Atlantic cover a widespread area over the continent and persist for a longer period. mT air masses from the Gulf-Caribbean-Atlantic source region also receive a lot of rainfall in the eastern two-thirds of the United States. Each of these types of air masses has some specific features associate with temperature and moisture distribution and air stability.

16.3 Air Mass Modification

As the air masses move away from the source region, can become modified. In its source region, an air mass gains properties which are characteristic of the underlying surface. It may be cold or warm, and it may be dry or moist. Modifications of air masses may arise from:

(a) Temperature differences between air mass and surface
(b) Vertical movements induced by cyclones and anticyclones
(c) Topography of the terrain

An air mass moving over the sea is said to have a maritime track. This air mass will typically increase its moisture content, particularly in its lowest layers, by evaporation of water from the sea surface (as seen in the above diagram). On the other hand, an air mass moving over the land (with a continental track) will remain relatively dry. A cold air mass flowing away from its source region over a warmer surface will be warmed from below making the air more unstable in its lowest layers. A warm air mass flowing over a colder surface is cooled from below and becomes stable in its lowest layers.

16.4 Fronts (Frontal Boundaries)

Frontal zones (fronts) form between a two contrasting air masses. They represent boundaries surfaces which separate air masses of different densities. Weather fronts mark the boundary or transition zone between two different air masses, which often have contrasting properties (Carlson 1991; Spiridonov and Ćurić 2011; Bott 2012; Ćurić and Janc 2016). For example, one air mass may be cold and dry, and the other air mass may be relatively warm and moist. Weather fronts mark the boundary or transition zone between two air masses and have an important impact upon the weather. One air mass is generally warmer and wetter than the other.

16.4.1 Types of Weather Fronts

According to the characteristics of air masses that occur at border areas, there are four (4) basic types of fronts:

- Warm
- Cold
- Stationary
- Occlusion

In the atmosphere of the border area between two air masses, other forms of fronts are manifested, such as the line of instability and vertical movements induced by cyclones and anticyclones or topography of the terrain.

Warm front occurs when the position of the front of the country is so warm that the air takes the territory previously covered by the colder air (Fig. 16.2). The presence of a warm front means that warm air is advancing and rising over cold air. This is because warm air is "lighter" or less dense than cold air. Warm air is replacing cooler air at the surface.

Cold front occurs where cold continental polar air actively progresses in the region, which is occupied by hot air (Fig. 16.3). The presence of a cold front means that cold air is advancing and pushing underneath warmer air. This is because the cold air is "heavier", or denser, than the warm air. Cold air is thus replacing warm air at the surface.

Stationary front occurs when air flow from both sides of the front is not the cold air mass or the warm air mass (Fig. 16.4).

Occluded front (see Fig. 16.5) develops when an active cold front is overtaking a warm front and gradually presses the top.

These are slightly more complex than cold or warm fronts. The word "occluded" means "hidden", and an occlusion occurs when the cold front "catches up" with the warm front. The warm air is then lifted from the surface and therefore "hidden". An occlusion can be thought of as having the characteristics of both warm and cold fronts. The main features of surface fronts and their respective characteristics are illustrated in (Fig. 16.6).

Fig. 16.2 Warm front

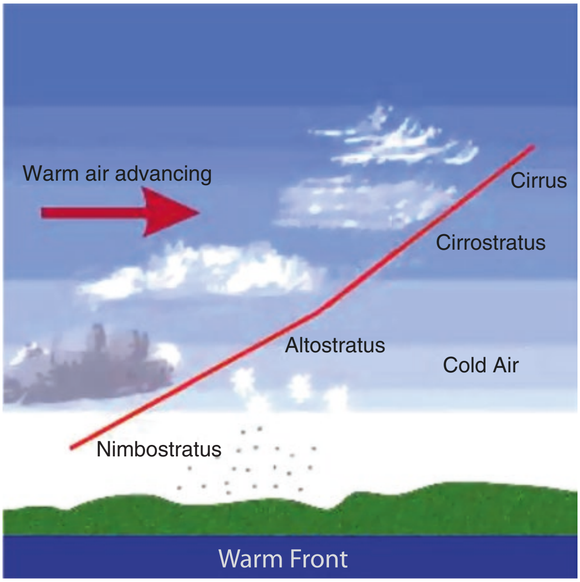
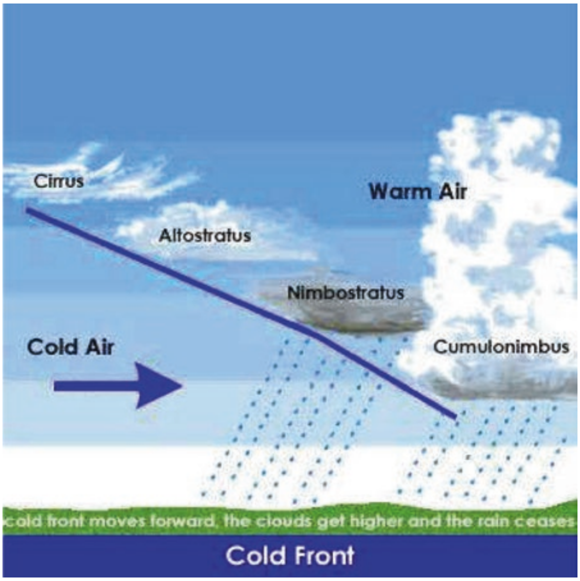


Fig. 16.3 Cold front



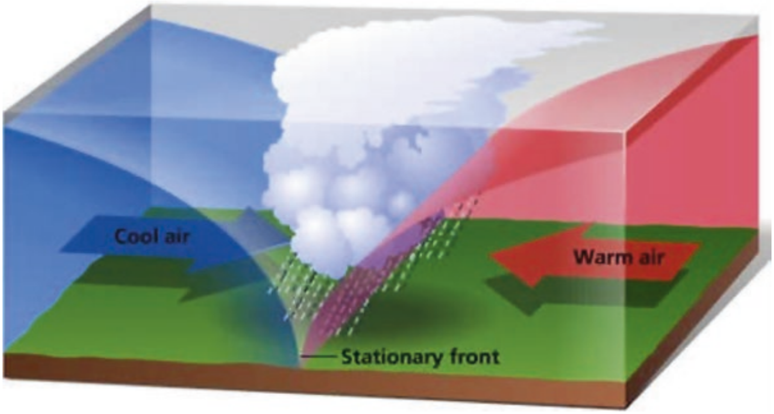


Fig. 16.4 Stationary front

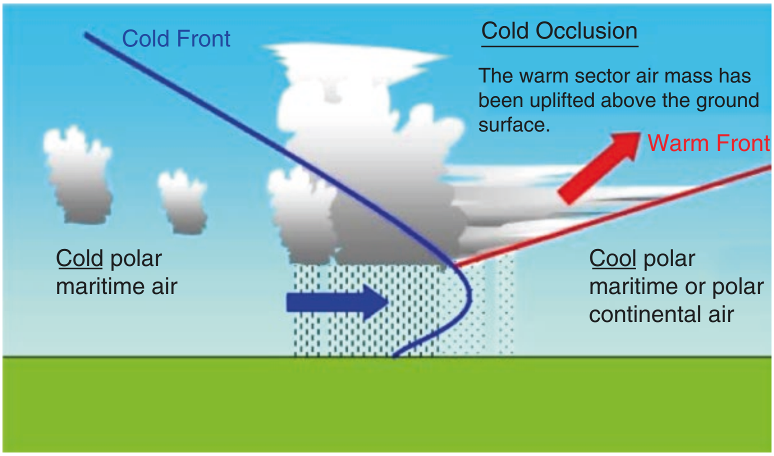


Fig. 16.5 Occluded front. Credit: Pierre_cbderivative work, https://commons.wikimedia.org/w/index.php?curid=27925717

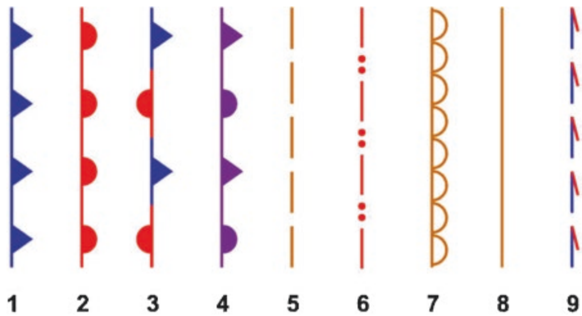


Fig. 16.6 Weather fronts. (1) Cold Front; (2) warm front; (3) stationary front; (4) occluded front; (5) surface trough; (6) squall line; (7) dry line; (8) tropical wave; (9) trowel – trough of warm air aloft

16.4.2 Occluded Fronts

There are two types of occluded fronts:

1. Cold-type occluded front, where the air behind the cold front is colder than the fresh air that it is overtaking
2. Warm-type occluded front, where the air behind the cold front that is moving forward is warmer than the cold air that is overtaking

At the moment when they formed "the wave", the warm air is advancing towards the pole, taking the area previously affected by colder air.

This change in direction of flow causes surface modification (re-adjustment) in the field of pressure, resulting in almost circular isobar, low pressure placed on top of the wave. Usually, the position of the cold front moves faster than that of the warm front, and the warm front begins to rise.

This process is known as occlusion forming occluded front. In addition to these basic forms of atmospheric fronts, there are also other types or transient forms (e.g. height warm front, height cold front, squall lines, trough line, convergence line, wedge line, etc.).

The instability line or a squall line represents a linearly oriented transition zone between the dry, dense air and less dense, moist, often associated with the occurrence of severe thunderstorms (potential tornadoes), especially during a warm season.

*Upper warm front***.** A high-altitude warming represents the same physical process as a warm front; however, the air mass change takes place only in height. The passage of an upper warm front may bring warmer air at an altitude of 3,000 m, without front signals on the ground such as an increase in temperature or bringing a change of air mass at the surface.

Upper cold front. A vertical cold front represents the same physical process as a cold front, but the air mass change takes place only in height. So, there are no front signals on the ground, such as a decrease in temperature to observe, clouds and rain are of course also from the ground out.

Trough line. A trough line marks the trough of a pressure wave in higher layers (e.g. 500 hPa levels). In this area, an advance of cold air takes place in the height; on the other hand, the air is stabilized at the front by dynamic lifting. This often results in rain or snow showers and thunderstorms.

Wedge line (wedge of relative topography). The relative topography is a measure of the energy content of an air mass. With a wedge in the relative topography, the most energetic area within an air mass is found. In summer, thunderstorms usually develop on these lines.

Convergence line. A convergence line describes the air convergence at the near surface layer. Basically, the convergence line is a narrow band of cloud that remains stationary and can produce large amounts of rain across a relatively small area. This phenomenon can be identified by the surface wind. When air masses flow together, they are forced to ascend, which, given enough humidity, is accompanied by cloud formation and, subsequently, precipitation formation. In summer, thunderstorms are often triggered along convergence lines.

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Chapter 17
Cyclones and Anticyclones

17.1 General Terms

The terms cyclone and anticyclone are used to describe organized atmospheric systems with circular flow areas with low and high atmospheric pressure (e.g. Martin 2006; Russell and Thompson 2002; Wallace and Hobbs 2006; Holton 2004). In the Northern Hemisphere, wind in the low-pressure systems (cyclones) is blowing counterclockwise, while anticyclones spin clockwise in the Northern Hemisphere and counterclockwise in the Southern. The main structure and the circulation of the cyclone and anticyclone are shown on Fig. 17.1. In the case of the cyclones, the pressure is lowest the centre (depression), while in the case of the anticyclones, the pressure is highest at the centre. In the first case, the pressure gradient is directed towards the centre, while in the latter case, the pressure gradient force is directed from the centre of high pressure to the periphery. When cyclone or anticyclone is associated with the wave front, then it is called a wave or frontal storm or mid-latitude anticyclone. Cyclones are associated with unstable atmospheric conditions and vertical movements of air. In cyclones, the air near the Earth moves towards the inside, the centre of the cyclone, where pressure is lowest, and then it starts rising. At the same height, the lifted air starts to diverge to the surrounding environment, away from the centre of cyclone.

One of the key factors which determine the formation of cyclones and anticyclones may be the development of irregularities in jet stream (Carlson 2012). When jet streams in the upper atmosphere start to oscillate back and forth along the east-west axis, they can affect cyclone or anticyclone systems that already exist in the lower troposphere. As a result, a relatively stable cyclones (or anticyclones) may develop, and they travel in the eastern or north-eastern direction through the continent. In relatively rare circumstances, such systems can gain enough energy, which can be destructive to human life and infrastructure. Storms, tornadoes, and vigorous hurricanes are examples of such extreme conditions.

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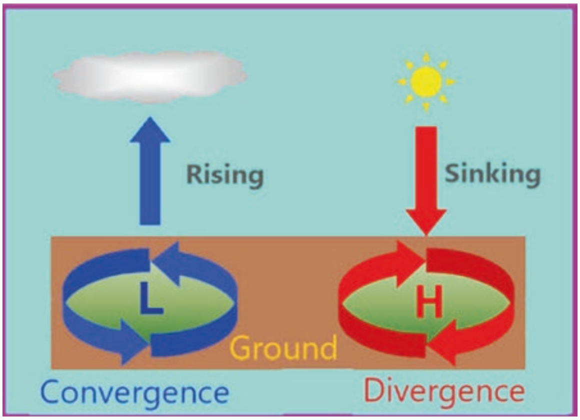


Fig. 17.1 Low- and high-pressure systems

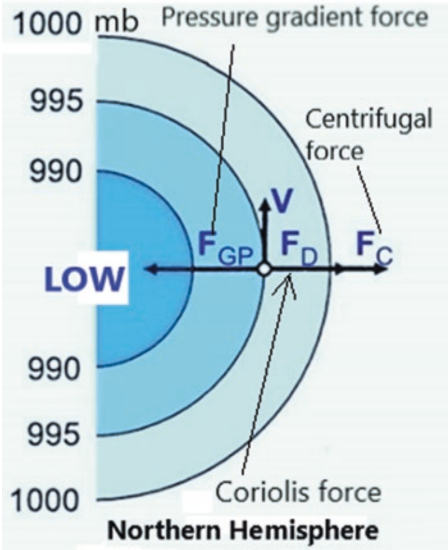


Fig. 17.2 Structure of the flow field of the cyclone

17.2 Low-Pressure Systems: Cyclones

A low organized pressure system, also known as a depression, occurs when the weather is dominated by unstable conditions. Under a depression air is rising, forming an area of low pressure at the surface. This rising air cools and condenses and helps encourage cloud formation, so the weather is often cloudy and wet. In the Northern Hemisphere, winds blow in anticlockwise direction around a depression. Isobars are normally closely spaced around a depression leading to strong winds.

Depressions can be identified on weather charts as an area of closely spaced isobars, often in a roughly circular shape. If isobars are along the centre of the lowest pressure, they are called troughs. Cyclones are often accompanied by fronts. As the air enters the area with low pressure from all directions, as the result of the Coriolis force, the cyclonic flow is deflected towards the right (Fig. 17.2). This creates a counterclockwise rotation around the centre of low air pressure, while causing convergence near the centre of the system. Convergence is occurring in the bottom area of low pressure. Near the Earth's surface, friction plays a major role in redistribution of air inside the atmosphere, by changing the direction of its flow. As a result, movement of air angle occurs, which is normal with isobars, the area with lower pressure. Therefore, resultant wind is blowing counterclockwise to the surface cyclone in the Northern Hemisphere. Convergence occurs when the air coming from different directions moves towards individual location. As the air collides near the centre, it is forced to rise, and thus, by divergence air is removed from the centre of the system. Surface winds around the Northern Hemisphere cyclone (low air pressure) blow counterclockwise and converge. In that stage the wet air rises.

17.2.1 Formation of Extratropical Cyclone

Extratropical cyclones develop in an old stationary front, i.e. on the polar front where large horizontal temperature gradients with a forcing air lifting exist. In this stage there is convergence in cyclone, which causes air rising from the Earth's surface (LaBeau 2009). The second stage begins with the formation of instability. Warm air moves towards the northeast, while cold air moves towards the southwest. In this way the fronts are formed. Warm front moves northeast and cold front moves southeast. The region between the fronts is called warm sector. The low pressure continues deepening. A widespread rainfall appears in front of the warm front and a narrow band of snow at the cold front. Consequently, wind speed successively increases. The development stage takes place for about 12 to 24 hours. The warm front moves northeast. The region between fronts is called warm sector. At this stage low pressure continues to decline, and widespread rain occurs in front of the warm sector and a narrow band of snow at the cold front. Wind speed increases.

17.2.2 Life Cycle of the Extratropical Cyclone

Cyclones formed at high latitudes, near the polar front, take the form of a developmental wave (Byers 1944; Bjerknes and Holmboe 1944). These cyclones are called extratropical cyclones (Fig. 17.3).

The theory of cyclogenesis (formation of cyclones) was first developed by Norwegian meteorologists – Bjerknes, Solberg, and Bergeron. In 1922, well before routine upper air observations began, they transferred experience from analysing surface weather maps over Europe into the Norwegian cyclone model, a conceptual picture of the evolution of an extratropical cyclone and associated frontal zones at ground. They noted that the strongest temperature gradients usually occur at the warm edge of the frontal zone, which they called the front.

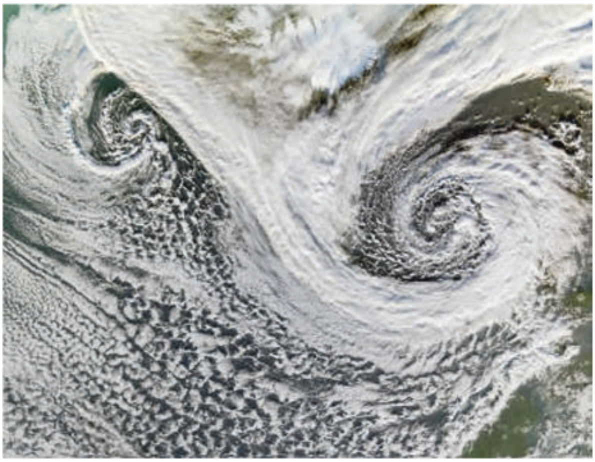


Fig. 17.3 Extratropical cyclones

According to the Norwegian model of the polar front, cyclones are formed along fronts and continue through a generally anticipated lifecycle.

The two air masses that have different densities are moving parallel to the front but in opposite directions. In that stage, cyclogeneses occur, as well as the formation of the frontal surface with the wave form, which is several hundred kilometres long (Bishop and Thorpe 1994). According to the theory of polar front, the front between cold polar air and warm sea air naturally creates wave shape form of the initial cyclone. The open wave with well-defined warm and cold front is deepening. Cold front then advances faster than the warm front, and it generates the occluded front. The occluded front is finally completed (Fig. 17.4). The lifecycle of extratropical cyclone or cyclone in high latitude is based on the school model of Bergeron. A more detailed illustration of all stadiums of the life cycle of the cyclone is given in (Fig. 17.5). According to this model, several successive stages in the life cycle of the storm or extratropical cyclone, which occurs in high latitudes, are separated:

- Frontogenesis (frontal wave)
- Mature stage (open wave)
- Developing stage (occluded storm)
- Dissipation of the cyclone

According to the values of sea level pressure in the centre of the cyclones in its earliest stage of development, it is evident that the pressure is reduced (990–1000mb). The maximum enhancement occurs in the developed stage of the cyclone (960–990 hPa), the final phase of decay slowly growing in the interval (960–990) to 1010 hPa.

There are three types of cyclones:

- Wave cyclone (occurs in high geographic latitudes, Arctic, Antarctic zones)
- tropical storm (it occurs in tropical and subtropical zones)
- Tornado

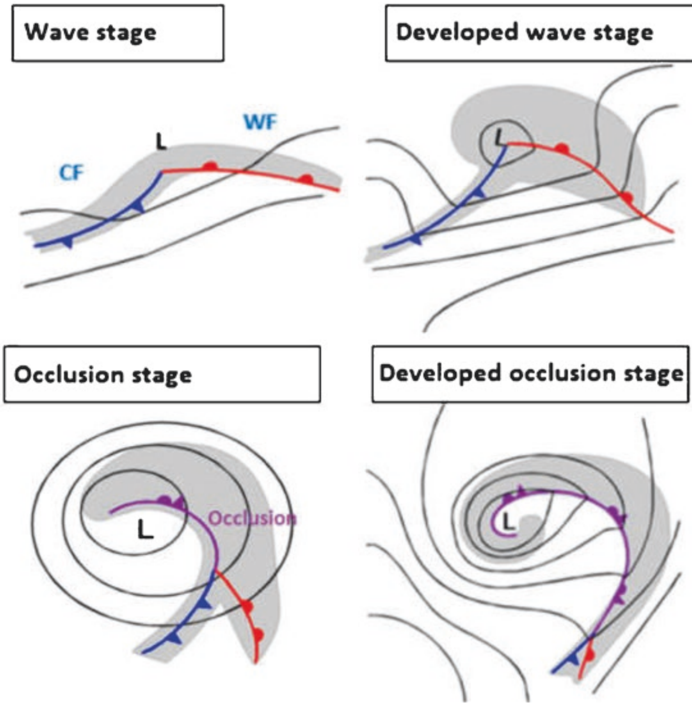


Fig. 17.4 Four stages of the classical polar front theory (Credit: http://www.eumetrain.org/satmanu/polar_front_theory/index.html)

A wave cyclone is a cyclone that travels at high latitudes and includes interaction of cold and warm air masses along the tightly defined fronts. A typical example of wave cyclone is shown in (Fig. 17.6).

17.2.3 Cyclone Movement

A cyclone usually moves towards east (or north-east). Driven by the western winds aloft, cyclones usually move through Eastern Europe. When idealized wave storm moves over the region through which the warm front sets the area under the influence of tropical maritime air mass (mT), which is usually characterized by high temperatures, there are southern winds and clear skies. Passage of cold front is easily detected by the shift of wind, the replacement of air over southern or southwest winds with wind from west or northwest. Also, there is a significant fall in temperature. As shown by the model of the polar front, cyclogenesis occurs where the frontal area is in a wave-shaped discontinuity. Flow of air at a height (divergence and convergence) plays an important role in maintaining and cyclonic and anticyclones circulation. Divergent field of cyclone aloft contributes air flow to be directed

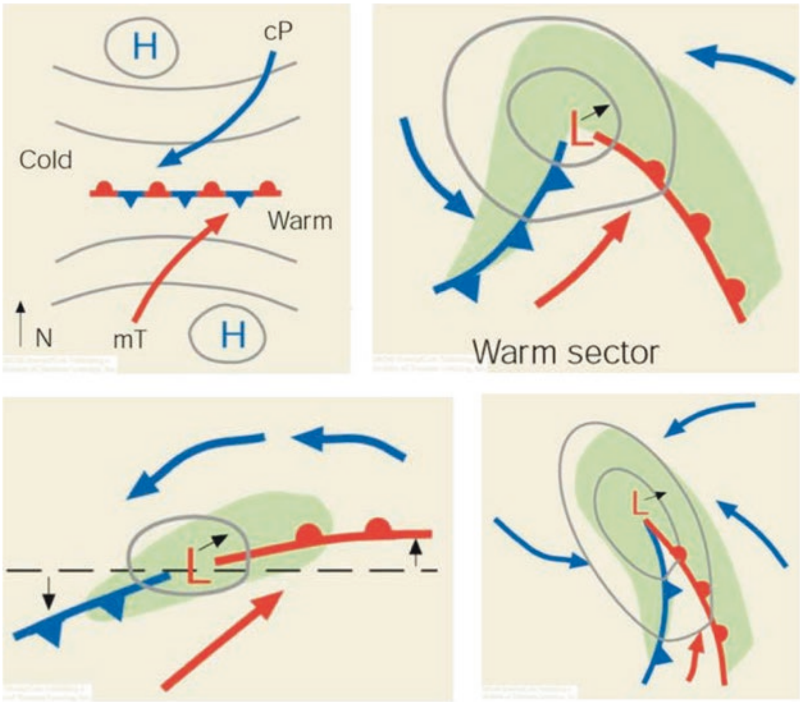
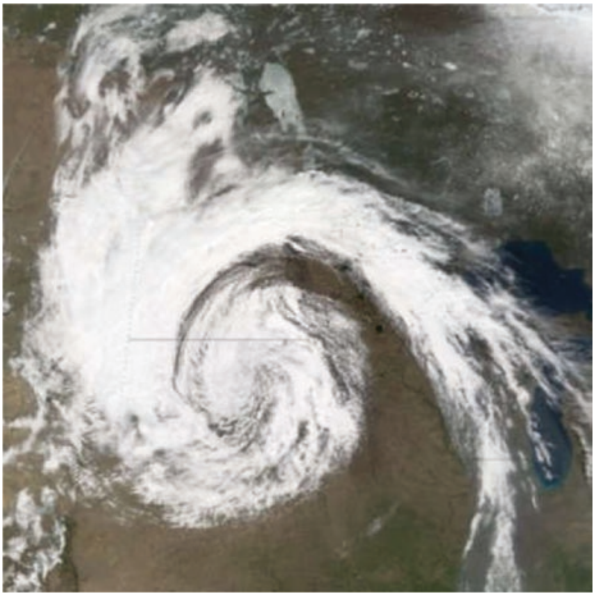


Fig. 17.5 Life cycle of extratropical cyclone, (Norwegian model) (Credit: https://slideplayer.com/slide/2550292/)

towards the outside in all directions. Usually, winds blowing from west to east are widespread along the curves. During colder months, when temperature gradients are expressed, cyclonic storms are frequent. Furthermore, the western air flow at a height that tends to manage the pool of bilge systems is usually developed in the direction west east. During the cold months, when temperature gradients are steep, cyclonic storms are more frequent. Western flow of air at a height aims to streamline these systems developed pressure in the general direction from west to east. In spring, stressed temperature gradient of the Earth from north to south can generate intense cyclonic storms. In mid-latitude areas when the storm passes into summer with its accompanying fronts, temperatures can change quickly, from non-seasonal warm to non-seasonal cold, and thunderstorms with hail may occur, even followed by snow. In the final phase, the warm sector of the cyclone redraws, and occlusion grows. All energy from the temperature difference is used for lifting the warm air. Then, the cold air descends, so that it comes to stabilize atmospheric conditions. Dissipation of the cyclone begins when almost all the warm sector moves up and cold air surrounds the cyclone at lower levels. At this point, the cyclone is exhausting its source of energy, and suddenly, cyclonic highly organized system of circulation ceases to exists.

Fig. 17.6 Wave cyclone



17.2.4 *Upper Level Low*

Upper Level Lows are closed cyclonically circulating eddies in the middle and upper troposphere. They are sometimes also called "cold drops", because the air within an Upper Level Low is colder than in its surroundings. The development of a typical Upper Level Low goes through four stages, during its evolution, starting from the initial formation of upper air level trough, which a bottom of an upper trough is separated from the main flow, until it finally fills up or merges with another trough (Fig. 17.7):

- Upper level trough
- Break-off
- Cut-off
- Final stage

1. Upper level trough stage. The main driving forces of the forming of the Upper Level Low are unstable waves within the mainstream, where the temperature wave is behind the geopotential wave. There is cold advection within the trough and warm advection on the ridge of the geopotential wave. The vertical axis of the trough has a backward-oriented inclination with height. The amplitudes of the waves increase; the wavelength can decrease.
2. Break-off stage. The amplitudes of the waves increase further. The isohypses form an inverse omega shape, and the cold air flows into the middle of this omega. Often at the same time, the ridge behind the main upper trough continues

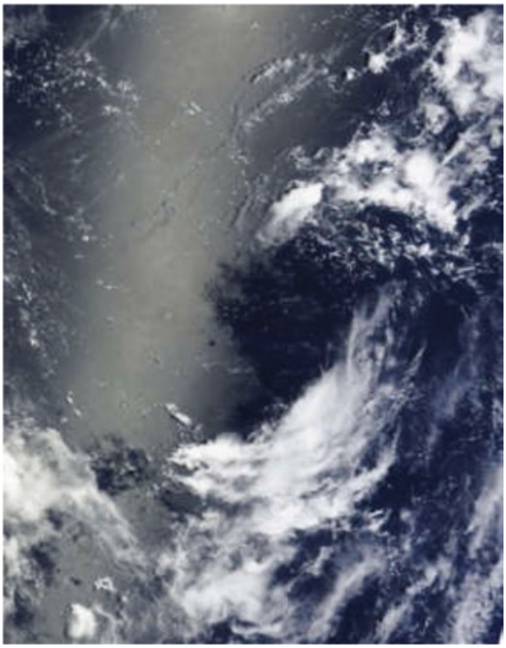


Fig. 17.7 Upper Level Low

to move eastwards faster than the trough, appearing to "fall forward". In the end of this stage, the cold bottom of the trough is separated from the mainstream.

3. Cut-off stage. The bottom of the upper trough is completely detached from the mainstream forming a closed circulation. If there is a strong forward-falling ridge behind, it may also separate from the mainstream and form an upper level high (a counterpart for the Upper Level Low). This happens in most of the ULL cases. The cold core of the Upper level Low warms up slowly because of the diabatic warming of the sinking clod air. If a cold Upper Level Low is situated over a warm surface, convection arises within the core. Another location for convection is ahead of the low within the area of a thickness ridge.
4. Final stage. Within an Upper Level Low, there is convection, unless the surface is very cold. The air near the surface is warm, and the circulation is slowed down by the friction. The convection brings warm air and friction upwards. Consequently, the Upper Level Low weakens slowly. In most cases the Upper Level Low merges with the mainstream before it has completely dissolved by the convection. Usually a large trough in the mainstream approaches from the rear and catches the upper level low. The Upper Level Low can also merge with another Upper Level Low.

17.3 High-Pressure System: Anticyclone

A high-pressure system with organized-closed circulation, also known as an anticyclone, occurs when the weather is dominated by stable conditions. Under an anticyclone air is descending, forming an area of higher pressure at the surface. Because of these stable conditions, cloud formation is inhibited, so the weather is usually settled with only small amounts of cloud cover. In the Northern Hemisphere, winds blow in a clockwise direction around an anticyclone. As isobars are normally widely spaced around an anticyclone, winds are often quite light.

Anticyclones can be identified on weather charts as an often-large area of widely spaced isobars, where pressure is higher than surrounding areas. Anticyclone is a system with high air pressure in the centre, which causes a stable, quiet, and fair weather. Isobars are nearly circular, and if found along the centre with the highest pressure, they are called ridges. In areas with high pressure, air is found on the surface as a result of convergence, which occurs in the upper layers. As air approaches the surface, it diverges from the centre (Fig. 17.8). Coriolis effect deflects the air to the right, and its path creates rotation around the centre of the high air pressure clockwise.

Divergence happens at the centre of the system with high air pressure on the Earth's surface. Divergence occurs when air is directed towards the outside of a specific location in different directions. Wind divergence in the high layers of the system held like the area with low air pressure. Without the presence of divergence, the system would be filled with air, and horizontal differences in pressure would be equalled, causing it to disappear. Figure 17.9 shows a high-pressure system (anticyclone). The high surface pressure area, with a high of about 1031 hPa, and clear clouds are well evident from the satellite image.

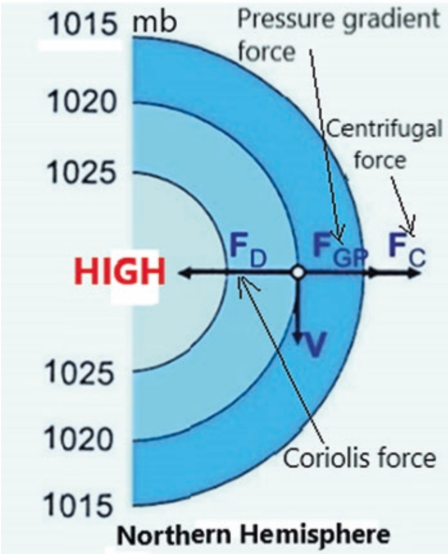


Fig. 17.8 Anticyclone structure

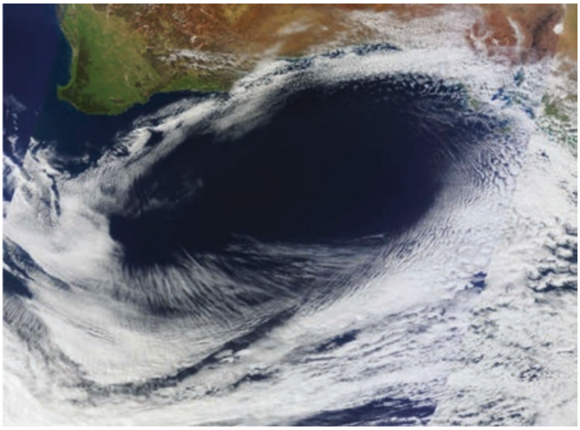


Fig. 17.9 High-pressure area seen from the satellite

Unlike cyclones, which tend to move towards the northeast, anticyclones often move southwest. As a result of the gradual pressure increase towards the centre, anticyclones cause clear skies and stable and calm weather conditions. Large systems with high pressure, called blocking anticyclones, are found in high latitudes each winter, and they approximately distance themselves with zonal flow from west to east and go towards the pole. Stagnant anticyclones move eastward, keeping one part dry naturally in the duration of 1 or more weeks, while the other region remains constantly under the influence of cyclonic storms. Also, due to the air subsidence, large stagnant anticyclones can create temperature inversion, which contributes to increasing the level of air pollution. In winter the clear, settled conditions and light winds associated with anticyclones can lead to frost and fog.

The clear skies allow heat to be lost from the surface of the Earth by radiation, allowing temperatures to fall steadily overnight, leading to air or ground frosts. Light winds along with falling temperatures can encourage fog to form; this can linger well into the following morning and be slow to clear. In summer the clear settled conditions associated with anticyclones can bring long sunny days and warm temperatures. The weather is normally dry, although, occasionally, very hot temperatures can trigger thunderstorms.

17.4 Weather Conditions Associate with Cyclones and Anticyclones

Weather in the area of baric systems depends on their dynamic structure and thermodynamics. Different forms of weather tend to be associated with cyclones and anticyclones. The characteristic signs of cyclones or low-pressure systems are unstable and variable weather conditions. The weather is cloudier, wetter, windier, and occurrence of rainy, stormy, and unfair weather. The anticyclone systems or high air pressure systems (highs) are mainly associated with appearance of stable, stagnant, calm, and fair weather conditions.

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Chapter 18
Tropical Storms

18.1 General Overview

Tropical cyclones represent the most severe storms (Longshore 2008a, b). By definition, tropical storm is a system with low air pressure in the centre and organized circulation (Fig. 18.1) which develops over tropical and subtropical ocean (Krishnamurti et al. 2013; Allaby 2014; Anthes 2016). This term encompasses tropical depressions, tropical storms, hurricanes, and typhoons. At mature stage, the tropical cyclone is one of the most intense and tropical storms, with winds which exceed 90 m s^{-1} and torrential rains. Its dimension usually ranges in diameter from 100 to well over 1000 km in a fully mature stage up to 2000 km in developing stage.

A tropical cyclone is characterized by a warm core in the centre, very steep pressure gradients, and strong spiral cyclonic inward winds near the Earth's surface becoming more nearly circular near the centre (Terry 2007). Pressure gradients, and resulting winds, are nearly always much stronger than those of extratropical storms. The cloud and rain patterns vary from storm to storm, but in general there are spiral bands in the outer vortex, while the most intense rain and winds occur in the eyewall. Depending on the speed of wind, tropical cyclones are divided into:

- **Tropical depression** (tropical cyclones whose maximum wind speed is less than 60 km/h)
- **Tropical storms** (when the maximum wind speed ranks between 60 and 110 km/h
- **Tropical cyclones** (when the maximum wind speed exceeds 110 km/h)

In the North and the North Pacific region, tropical cyclones are called "hurricanes" (Fig. 18.2); in the Western North Pacific region, they are named as "typhoons", while in the region of the Indian Ocean, they are known as cyclones. Figure 18.3 shows a sequence of typhoon Soulik near Japan, on August 20, 2018, using the Moderate Resolution Imaging Spectroradiometer (MODIS) on NASA's Terra satellite.

V. Spiridonov, M. Ćurić, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_18

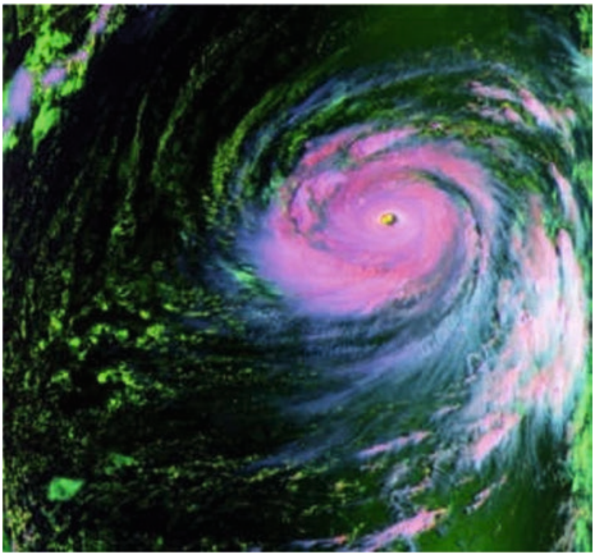

Fig. 18.1 Tropical storm

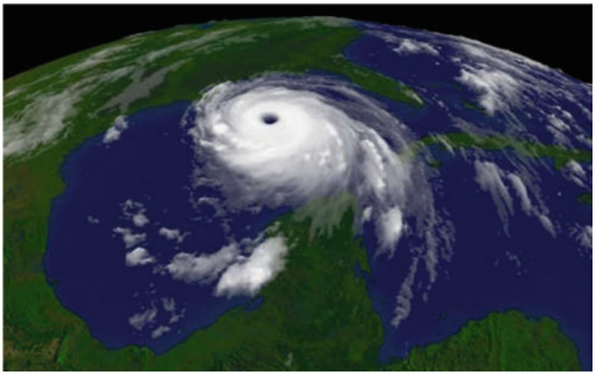

Fig. 18.2 Hurricane Katrina (Courtesy: NOAA/NESDIS: http://www.nnvl.noaa.gov/)

18.2 Tropical Cyclone Formation

The formation of tropical cyclones is still a topic of intense research and is not fully understood, although research has shown that certain factors must be present for cyclones to intensify to hurricane strength. The process of disturbance formation in the field of pressure and subsequent strengthening of the tropical cyclone depends on the following conditions:

1. Contact of the tropical storm in the warm ocean waters, where the phase of development is. This usually occurs between 5 and 20 deg. north and south latitude.

Fig. 18.3 Satellite image of Typhoon Soulik (Source: https://earthobservatory.nasa.gov/images/92628/typhoon-soulik)

2. Flow of moisture through evaporation of warm ocean waters near the equator, as significant source of latent heat energy used for intensification of the tropical cyclone.
3. Shape of the wind flow near the ocean surface, which forces the air vorticity inside.

As a result of these processes, a vertical column of air occurs, with very low pressure at its centre. In this form, groups of storms are allowing air to be further heated and raise high into the atmosphere.

18.2.1 General Factors

Basic factors for the formation of tropical cyclones are atmospheric disturbances that occur as a result of wave movements, atmospheric disturbances (instability lines or squall lines), thermobaric systems, and frontal boundaries (Oouchi and Fudeyasu 2013). Some of them are listed and described below:

- **Eastern waves** (tropical waves): waves that are turned into valleys with low air pressure, they move westwards in the tropical belt of the eastern winds (eastern winds). The rough is a region with relatively low pressure. Much of the tropical cyclones are formed by the waves of the eastern winds.
- **West Pacific line disturbances**: represent a line of convection (like a line of instability), which is formed over the West African continent and is moving into the Atlantic Ocean, usually faster than tropical waves.
- **Tropical-upper troposphere upper trough**: a trough or cold core low air pressure (depression) in the upper atmosphere, which creates favourable conditions for convection and developing tropical storm;
- **Old frontal boundary** occurs in the remnants of the polar front, where lines of convection as precursors to the formation of tropical storm are created. In the

Atlantic Ocean storms, it occurs in early or late season hurricanes in the Gulf of Mexico or the Caribbean Sea.

18.2.2 The Basic Ingredients

The basic ingredients which serve as a triggering mechanisms for initiation of formation of tropical cyclones (hurricanes, typhoons, tropical depressions, tropical storms) are the sea surface temperature (SST) distribution, the geographical location, some pre-existing mesoscale atmospheric disturbance, prevailing winds, thermodynamic conditions, environmental instability parameters, and vertical stratification of the atmosphere (Longshore 2008a, b; Galvin 2015). It is well accepted that interannual perturbations in sea surface temperature (SST) in the ocean waters (Atlantic Ocean) are associated with variations in seasonal tropical cyclone frequency. Sea surface temperature distribution is an important factor as it represents an important source of latent heat release through evaporation of warm ocean waters. Therefore, the formation of tropical cyclone requires sufficiently warm sea surface temperatures (SSTs) usually greater than 26 °C. Because SST are seasonally changing, thus the likely locations of tropical storms formation also change. However, the most convenient areas are over tropical ocean waters away from the equator. Other significant ingredient is a pre-existing low-level atmospheric disturbance that could be generated as the result of energy transfer from microscale to mesoscale or some other scale interaction processes such as generation of internal mesoscale instabilities or interactions of cloud physical processes with mesoscale. Other factors include a high mid-level moisture, weak vertical wind shear.

18.3 Areas of Formation

Considering that the temperature of sea level should be at least (27 °C) for tropical cyclogenesis, it is natural that they are formed near the equator. However, only in rare circumstances, tropical cyclones form at distance 5 degree latitude from the equator.

1. Atlantic basin (including the North Atlantic Ocean, Gulf of Mexico, and Caribbean Sea)
2. North-eastern Pacific basin
3. Pacific Northwest basin, including the South China Sea
4. North Indian basin (including the Gulf of Bengal and Arabian Sea)
5. South-western Indian basin from Africa to about 100 E
6. Southeast Indian/Australian basin (100 ° E to 142 ° E)
7. Australian/Southwest Pacific basin (from 142 ° E to 120 ° W)

Fig. 18.4 Regions of formation of tropical cyclones (Credit: http://basins.ghkates.com/tropical-cyclone-basins/)

This is due to the weak influence of Coriolis force, which causes cyclone fluctuations. There are seven standard basins where tropical cyclones form (Fig. 18.4). Table 18.1 provides a brief description of the area of formation, the main characteristics, and the period of activity of tropical cyclones.

18.4 Classification of Tropical Cyclones

Tropical cyclones with organized low-pressure systems accompanied by thunderstorms, with a clearly defined circulation and permanent winds of about 60 km/h or less, are called "tropical depressions". In case when cyclonic circulation becomes more organized and maximum sustained winds gust between 60 and 120 km/h, it is commonly called "tropical storms". If persistent winds reach a maximum speed of about (120 km/h), tropical cyclones are named as follows:

- **Hurricane** (North Atlantic Ocean, northeast Pacific Ocean, and southern Pacific Ocean east of 160 degree E). The word "hurricane" comes from the Caribbean Indians of Western India, who called the Storm "huracan". It is assumed that Taint's ancient tribe in Central America called their god of evil "Huracan". Spanish colonists modified the word to "hurricane".
- **Typhoon** (Northwest Pacific). Super typhoon occurs if maximum winds are constant at least 241 km/h.
- **Tropical storm** (southwest Pacific, west of 160 °E, and the southeast Indian Ocean, east of 90 ° E).
- **Storm cyclone** (in the northern Indian Ocean).
- **Tropical storm** in the south-western Indian Ocean.

Table 18.1 Basic features of tropical cyclones

Atlantic basin, which includes the North Atlantic Ocean, Gulf of Mexico, and Caribbean Sea	The hurricane season officially starts on June 1 and lasts until November 30. Weak activity since mid-September
North-eastern Pacific basin	Wide peak activities, which began in late May or early June and last until late October or early November, with a maximum intensity of storms in late August
North-western Pacific basin, including the South China Sea	It happens regularly throughout the year, with a minimum intensity in February and the first half of March. The main season lasts from July to November, with a maximum intensity of tropical cyclones in late August
North Indian basin (including the bay of Bengal and Arabian Sea)	Double maximum intensity (peak) of the activity of tropical cyclones is noticed in the period (May–November) and (April–December). Powerful cyclones with storm winds (> 119 km / h) usually occur in (April–June) and again in late September until early December
South-western Indian basin (from Africa to about 100 E)	Begins in late October (early November). Achieves double peak in activity, one in the middle of January and another in mid-February and lasts until the beginning of March and then ends in May
South-eastern Indian-Australian basin (100 ° E to 142 ° E)	Begins in late October (early November). Achieves double peak procedure, one in the middle of January and another in mid-February and lasts until the beginning of March, and it ends in May
	A delay in activities of the Australian/Southwest Indian basin in February is more emphasized than the delay of the south-western Indian basin
Australian-Southwest Pacific basin. From 142 ° E to about 120 ° W	Begins in late October (early November). Single peak reaches the end in February and then weakens in early May

18.5 Tropical Storm Structure

The key parts of the rainy tropical storm tracks are eye and eye layer (eyewall) as it is shown in Fig. 18.5. The air rotates towards the centre of the cyclone counterclockwise in the Northern Hemisphere (and clockwise in the Southern Hemisphere) and then comes on the top in the opposite direction.

At the centre of the storm, air descends and forms an “eye”, which is usually not accompanied by the appearance of clouds. The air rotates towards the centre of the cyclone counterclockwise in the Northern Hemisphere (and clockwise in the Southern Hemisphere), and then comes on the top in the opposite direction.

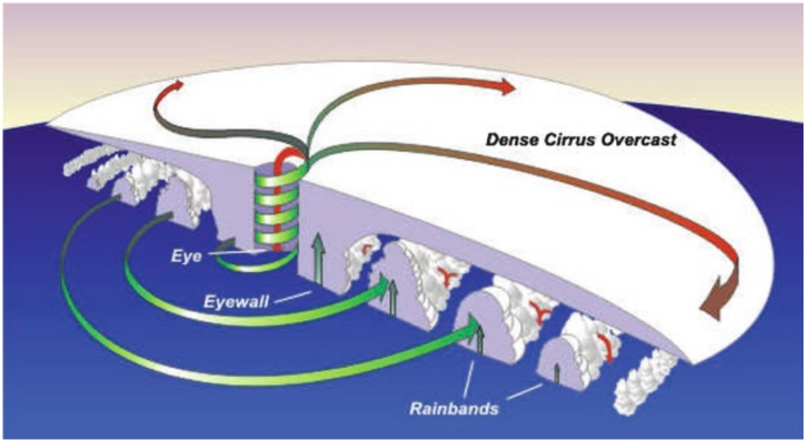

Fig. 18.5 Structure of tropical storm

18.5.1 Cyclonic Eye

Cyclonic eye is found at the centre of strong tropical cyclone. It is usually circular and typically 40–65 km in diameter. It is surrounded by the vertical wall of thunderstorms (*eyewall*), where the most severe weather of a cyclone occurs. The eye is possibly the most recognizable feature of tropical cyclones. In strong tropical cyclones, the eye is characterized by light winds and bright, clear skies, and warmer weather condition. In weaker tropical cyclones, the eye is less well-defined and can be covered by the central dense overcast, which is an area of high, thick clouds which show up brightly on satellite pictures. Weaker or disorganized storms may also feature an eyewall which does not completely encircle the eye or have an eye which features heavy rain. In all storms, however, the eye is the location of the storm's minimum barometric pressure: the area where the atmospheric pressure at sea level is the lowest. Near the sea surface, air temperature is almost uniformed throughout the storm. It is assumed that the eye of the tropical storm is formed with a combination of maintaining angle moment and centrifugal force. In this case, the air increases its speed when it points towards the centre of the tropical storm. With increasing the speed, strength appears to be directed towards the outside, and it is called centrifugal force. It occurs because of the angle moment that the wind tends to carry through the line. As the wind flows around the centre of the tropical storm, the force is removed to the outside. A more pronounced curve and/or faster rotation results in stronger centrifugal force. The strong rotation of air around the cyclone, with wind speeds of ~119 km/h, fits the air in the centre, forcing its rapid rising to the higher layers, at the height of about (16–32 km) from the centre, forming a wall eye. This strong rotation also creates a vacuum of air at the centre, causing part of the air flows to the top of the eye wall, to replace lost air mass near the centre. By this lowering, the air tends to create a clean air zone in the centre, without the presence of clouds. People often perceive the passage of the cyclonic eye at night, watching the stars in the sky. Sudden change with the emergence of very strong winds, which occur immediately after the silent condition, is dangerous to people who neglect the structure of tropical cyclone (hurricane). They usually notice the occurrence of weak wind and a fine weather in the eye, maybe for moment thinking that the hurricane has passed, but storm is only half the opposite side, with dangerous winds for a few minutes back in the wall of the eye, this time from the opposite direction.

18.5.2 Eyewall

The most dangerous and destructive part of a tropical cyclone is the eyewall shown on Fig. 18.6. Here winds are strongest, rainfall is heaviest, and deep convective clouds rise from close to Earth's surface to a height of 15 km. As noted above, the high winds are driven by rapid changes in atmospheric pressure near the eye, which creates a large pressure gradient force. Winds reach their greatest speed at an altitude of about 300 metres above the surface, while closer to the surface they are slowed by friction. Higher than 300 metres, they are weakened by a slowing down of the horizontal pressure gradient force as the results of temperature structure of the storm. Air is warmer in the core of a tropical cyclone, and this higher temperature causes atmospheric pressure in the centre to decrease at a slower rate with height than occurs in the surrounding atmosphere. The reduced contrast in atmospheric pressure with altitude causes the horizontal pressure gradient to weaken with height, which in turn results in a decrease in wind speed. Friction at the surface, in addition to lowering wind speeds, causes the wind to turn inwards towards the area of lowest pressure. Air flowing into the low-pressure eye cools by expansion and in turn extracts heat and water vapour from the sea surface. Areas of maximum heating have the strongest updrafts, and the eyewall exhibits the greatest

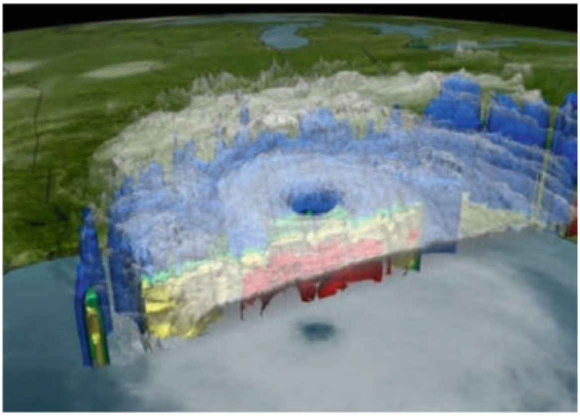

Fig. 18.6 Eyewall of tropical storm

vertical wind speeds in the storm – up to 5 to 10 m/s. While such velocities are much less than those of the horizontal winds, updrafts are vital to the existence of the towering convective clouds embedded in the eyewall. Much of the heavy rainfall associated with tropical cyclones comes from these clouds. The upward movement of air in the eyewall also causes the eye to be wider aloft than at the surface. As the air spirals upward, it conserves its angular momentum, which depends on the distance from the centre of the cyclone and on the wind speed around the centre. Since the wind speed decreases with height, the air must move farther from the centre of the storm as it rises. When updrafts reach the stable tropopause, the air flows outward. The Coriolis force deflects this outward flow, creating a broad anticyclonic circulation aloft. Therefore, horizontal circulation in the upper levels of a tropical cyclone is opposite to that near the surface.

18.5.3 Rain Bands

In addition to deep convective cells (compact regions of vertical air movement) surrounding the eye, there are often secondary cells arranged in bands around the centre. These bands, commonly called rain bands, spiral into the centre of the storm.

In some cases, the rain bands are stationary relative to the centre of the moving storm, and in other cases, they seem to rotate around the centre. The rotating cloud bands often are associated with an apparent wobbling of the storm track. If this happens as the tropical cyclone approaches a coastline, there may be large differences between the forecast landfall positions and actual landfall. Sometimes there are delays between the spiral belts, which lead to no occurrence of rain or strong wind. As the tropical cyclone makes landfall, surface friction increases, which in turn increases the convergence of air flow into the eyewall and the vertical motion of air occurring there. The increased convergence and rising of moisture-laden air is responsible for the torrential rains associated with tropical cyclones, which may be in excess of 250 mm in a 24-hour period. At times a storm may stall, allowing heavy rains to persist over an area for several days. In extreme cases, rainfall totals of 760 mm (30 inches) in a 5-day period have been reported.

18.5.4 Dimensions of the Tropical Cyclone

Tropical cyclones usually extend over large area from 100 to 2000 km in diameter, although they can significantly change their dimension. In principle, the size (diameter) is not a basic indication of the intensity of the hurricane. For example, Hurricane "Andrew" (1992), second strongest, devastating hurricane that hit the North American continent, and then hurricane "Katrina" in 2005 that was a relatively small hurricane. According to a record, the largest storms like "typhoon" occurred in the north-western Pacific on October 12, 1979. They were accompanied by strong storm winds that are dispersed 1087 km in radius. The smallest tropical cyclone is "Tracy", manifested with strong storm winds which are dispersed only 48 km in radius, which affected Darwin, Australia, on December 24, 1974. Hurricane destructive winds and rains cover a wide belt. According to data for the Atlantic basin, the strongest storm was Hurricane "Wilma" in 2005, with the pressure of 882 hPa at the centre of the tropical storm, with permanent winds for about 280 km/h.

18.5.5 Tropical Cyclone Life Cycle and Tracks

Most tropical cyclones last for a few days, though some may take a while significantly longer or disappear within 12 hours. Although strong winds blow inside the cyclones, the cyclones themselves move very slowly (about 5 m/s), and their speed increases in the mature phase to about 20 m/s. Tropical cyclones, when they reach land, are weak and disappear rapidly as the flow of energy from the warm water surface stops. Rarely do they occur in temperate latitudes already transform into extant cyclones. Typical tropical cyclone trajectories are given in Fig. 18.7. Although tropical cyclone occurrence regions are known, as well as the fact that they move mainly westward, their exact trajectories are different and often difficult to predict. Hurricane movement also depends on the winds normal to their trajectories. Thus, if it forms in the tropical region of the eastern Atlantic, it moves westwards with the trade winds. In the Gulf of Mexico region, it turns northwest and poses a potential threat to the east coast of America and the Gulf Coast.

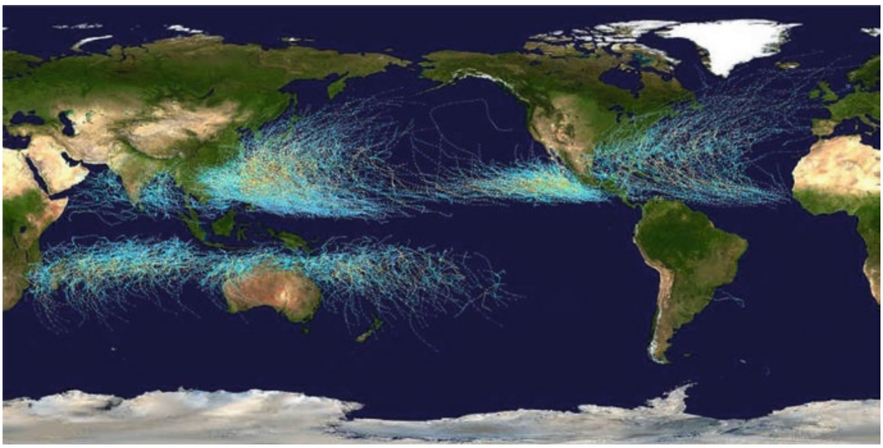

Fig. 18.7 Global Tropical Cyclone Tracks between 1985 and 2005

18.5.6 The Weather Associated with Tropical Cyclones

Tropical cyclones are always accompanied by heavy rain. Single storm can provide up to 3000 mm of rain. Intensive rainfall sometimes occurs several days after the overrun across the land and can also be very destructive and cause flooding. Due to steep pressure gradient, strong winds appear. The wind velocity rapidly increases from near zero in the eye of the cyclone to a maximum value at a radius between (10 and 100 km) from the bottom. The strongest winds occur near the head of storm.

18.5.7 Tropical Cyclone Disasters

Each year, beginning on June 1, Gulf countries and the east coast are in great danger of occurrence of tropical cyclones. Although many people know that the tropical cyclones occur with winds of destructive power, many hardly know that tropical cyclones directly or indirectly create other types of accidents that endanger life and material goods (Collins and Walsh 2019).

For that purpose, information and education about the formation, activities, and destructive power of tropical cyclones for better preparedness and taking preventive measures to reduce the risk of disasters are required.

18.5.8 Large Storm Wave

Destructive strength of tropical cyclones occurs not only due to the strong forcing wind but also as a result of the large storm wave and the accompanying waves it creates. Storm surge is water from the ocean that is pushed towards the shore by the force of the winds swirling around the hurricane (Fig. 18.8.) Storm surge combined with waves can cause extensive damage. It can severely erode beaches and coastal

Fig. 18.8 Hurricane storm wave

highways. Stormy wave is manifested by rapid sea level water rising (sometimes more than 6 metres). Many of the accidents associated with tropical cyclones appear due to the large storm wave. Large storm wave represents a water mass, which moves to shore, forced by the strength of the winds, which blow around the storm. This great leading wave is combined with ordinary waves, which create a hurricane storm wave.

The wave intensity is determined depending of slope of the continental shelf. Shallow slope of the shore allows flooding of coastal areas by the most powerful wave. Areas with steeper continental shelf reduce the likelihood of major flooding in the strong wave of hailstorms, although large destructive waves can still cause serious problems.

18.5.9 Windstorms and Flooding

Hurricanes are known by their destructive winds. Their intensity is measured by the rate of flow of the wind. The National Hurricane Centre issued a notification in respect of the wind and the category of hurricane.

This hurricane scale does not include wind shots or storms. Blows are brief, but with rapid breaks of the wind speed, which arises as a result of the disturbances. Cyclonic wind causes great damage to marine traffic and infrastructure on the mainland. Many homes were damaged and destroyed when strong wind simply raises the roofs of housing. The strong wind that moves over the top of the roof creates a lower pressure on its exposed side, the loft. Higher pressure in loft helps to cover rising. When loading, the roof part is blown away from the residence. When the roof is plucked, the walls are easier to waggle by the hurricane wind. Besides leading storm waves and strong winds, tropical cyclones are accompanied by strong rainfall and the occurrence of floods (Fig. 18.9). Even when the wind is turned down in several

Fig. 18.9 Tropical storms severe winds and flooding

subsequent days, there remains a potential threat of flooding from these storms. Tropical cyclones can also cause sudden flooding, urban flooding, and meat or river flooding. Sudden floods are weather situations that occur quickly. This type of flooding may begin within minutes or hours of heavy downpours. The rapid rise of water can reach up to 10 metres, and it can pull trees and destroy buildings and bridges. Urban floods are fast atmospheric phenomena; although not as intense as flash flooding, they may cause flooding down the street and flooding of basements and other structures. River floods are more lasting phenomena that occur when the expiration of torrential rains caused by the dissolution of hurricanes or tropical storms will reach the rivers. River flooding can happen in just a few hours, and it also lasts for a week or longer.

18.5.10 Names of Tropical Cyclones

Tropical cyclones have been given names, in order to provide easy communication between forecasters and the public, in terms of weather forecast, observations, and announcements. The first appropriate name for a tropical storm was used by Australian forecaster in the early twentieth century. He suggested that names of tropical cyclones are given “according to political figures”. During World War II, the tropical cyclones were given female names by the US Army and the Navy meteorologists (after their girlfriends or wives) who observed and gave forecast of tropical cyclones over the Pacific. From 1950 to 1952, tropical cyclones in the North Atlantic Ocean were identified by phonetic alphabet, but in 1953 female names were back. The World Meteorological Organization (WMO) and the US National Weather Service (NVS) progressed to list names that also included male names. The tropical cyclones in the Pacific Northwest region, officially from 1945, were given female names, and male names were also involved in the early 1979. From January 1, 2000, tropical cyclones in the north-western Pacific were named with new and very different list of names. The new names were Asian names from all nations and territories given by the Typhoon Committee of WMO. These newly selected names have two major differences from the rest of the world’s lists of names of tropical cyclones. Tropical cyclones have several male and female names, but the majority are names of flowers, animals, birds, trees, or even food. Finally, the names are not given in alphabetical order but arranged by the contributor in alphabetical order.

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Chapter 19
Thunderstorms, Lightning, and Tornadoes

19.1 Basic Definition

Storm is any disturbed state of the atmosphere, affecting the earth's surface and causing destructive weather. *In general storms represent* clouds with vertical development where the process of convection is dominant.

Convection is the vertical transport of heat and moisture in the atmosphere as *mode of the energy transfer*. According to another definition, convection is a thermally driven turbulent mixing in the atmosphere. There is a free convection, defined as a self-sustained flow driven by the presence of a temperature gradient (buoyancy force) or forced convection as mechanism in which fluid motion is generated by an external or mechanical factors (forced air lifting, horizontal mass convergence, cold air slides down over warm air, organized air circulation). We also distinguish a dry convection (as thermally driven-thermal lows) without clouds and precipitation and moist convection, where convective processes produce clouds and precipitation (Smith 1996; Pruppacher and Klettt 2010). In unorganized convection, there is a non-regular spatial distribution of cloud evolution generated by differential heating and air rising as less denser and warmer than environment. An organized convection is associated with occurrence of the regular spatial distribution of convective clouds evolution in a wider area. It usually occurs over surface boundary layer (1–2 km)-shallow convection, where Cu and Sc clouds are forming. In shallow convection exists a thermally driven turbulent mixing, where vertical lifting is capped below 500 hPa. Shallow convection is important for many aspects: cloud cover, vertical moisture transport, and surface precipitation in some regions and also as environment precursor for deeper modes of convection. In addition to shallow, also deep convection occurs that affects whole tropospheric layer, where thermally driven turbulent mixing and vertical lifting take place from lower atmosphere above 500 hPa. Triggering factors for their formation are frontal boundaries, shear lines, tropical cyclones, and tropical waves. They also develop under deep upward vertical motion (ahead of a polar front, organized tropical cyclone, positive scale Interaction, i.e.

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tropical wave moving under divergent side of upper trough/ridge) or during large scale air subsidence (behind Polar trough/ahead of block ridge, over continent generally inhibits convective development). General conditions for their formation are low-level convergence, upper level divergence, relative humidity in excess of 70%, unstable PBL layer. Storm initiation depends on the atmospheric instability and thermodynamic-environmental parameters as triggering factors for their initiation, evolution, and dissipation.

Thunderstorm or electrical storm is a local storm, produced by a cumulonimbus cloud and always accompanied by lightning and thunder, usually with strong gusts of wind, heavy rain, and sometimes with hail. It is usually of short life cycle hours for any one storm. A thunderstorm is a consequence of atmospheric instability. A strong convective updraft is a distinguishing feature of this storm in its early phases. A strong downdraft in a column of precipitation marks its dissipating stages.

19.2 Thunderstorm Formation

There are three important factors for the formation of thunderstorms: moisture, instability, and the mechanism of rising (Cotton et al. 2010). Typical sources of moisture are large water bodies such as the Atlantic and Pacific Ocean, Gulf of Mexico, Indian Ocean, and other large water areas. The sea surface temperature also plays an important role in controlling the source of moisture. Instability of the atmosphere occurs when air near the surface is heated or when the amount of cold air is moved to an area with a heated surface. In these situations, if the individual air particle is forced to move upwards, it will continue to rise independently by impact of the strength of displacement. As the air particle rises, so it cools, and some water vapour will condense, forming characteristic cumulonimbus cloud which represents thunderstorm. The basic features of storm are unstable air mass with heated moist air near the surface, with a cooler and dry air aloft. The air is forced upwards and will continue to rise, and the air that is driven downwards will continue to descend. For thunderstorm to occur, there should be a mechanism which initiates upper vertical movement. These are the main favourable conditions for lifting air: a) front-boundary between cold and warm air mass, as an ideal place for an outbreak of upward movements in the atmosphere as a place where instability occurs; b) dry-line boundaries between two air masses containing different amounts moisture; and c) constraints; if the wind blows over the obstacle, it will be forced to rise upwards.

19.3 Life Cycle of a Thunderstorm

Cumulus clouds appear before the formation of thunderstorms. Lifted cumulus cloud is associated by lifting of the air. Such lifting may be due to convergence of air associated with the anomaly in the heat on the Earth's surface or occurs as a result of inclined terrain. At this mature stage if rains occur, it would be quite a low rainfall, which would last about 10 minutes. The air continues to rise to the tropopause. The upper lid forces the air to spread out, resulting in the emergence of a form of “anvil”. Then, it is probably time for the occurrence of hail, heavy rain, often lightning, strong winds, and tornadoes. Storms are occasionally characterized by a dark green appearance. Duration is approximately 10 to 20 minutes, but some storms can last longer in the absence of wind shearing. This stage marks the disappearance of thunderstorms, and it is manifested by the characteristic descendent currents (downdrafts). Rainfall intensity decreases but still creates breakthroughs with strong descending movements near the surface, thus preventing the internal flow at lower levels. Lightning still represents a potential danger.

19.4 Thunderstorm Classification

Based on the number of primary convective cores, thermodynamic environmental conditions, and the microphysical structure (Pruppacher and Klettt 2004), storms can be classified into following groups: single cell, multicell, and supercell (Fig. 19.1). Formation of different types of storms is caused by different combinations of the above conditions.

(a) **Single cell.** In single cell storms, there is a core with vertical movement, called “cell”. Such a single cell is usually formed in isolation to other cells. Very often, more than one cell appears if the environment is suitable for development. They are typically driven by heating on a summer afternoon. Single cell storms produce hail, rain, and lightning. The lifetime is usually 20–30 min. Pulse storms can produce severe weather such as downbursts, hail, heavy rainfall, and weak tornadoes. Their evolution I spassing through three stages: cumulus, mature, and dissipation. The have weak wind shear and veering.

(b) **Multicell.** A group of cells mainly moving as a single units with each cell in a different stage of the thunderstorm life cycle. They have a multiple convective updrafts and downdrafts. According to their structure and evolutionary

Fig. 19.1 Storms classification

properties, they are classified as ordered, weak-ordered, and unordered multi-cell. They produce moderate hail, flash-floods, and weak tornadoes.

(c) **Supercell.** A special type of single cell storm that lasts longer than several hours. It has a very organized and strong vertical updrafts and downdrafts cores. Severe thunderstorms with rotating updrafts-mesocyclone (Ćurić 2001a; Spiridonov and Ćurić 2019). They are characterized with strong wind shear and veering, large CAPE, and storm helicity. In addition the directional wind shear (veering) at the near surface layer, high storm helicity index and differential heating induced by the strong local forcing environment serve as triggering factors for initiation of supercell storm with rotational updrafts-mesocyclone.

Supercell storms can produce strong downbursts, large hail, flash floods, and weak to violent tornadoes and can be very destructive phenomena. It is extremely dangerous and responsible for heavy rainfall, hailfall, flash floods, and even tornadoes. Thunderstorms are responsible for the development and formation of many severe weather phenomena. They are usually accompanied by strong winds and often produce heavy rainfall, hailfall, flash floods, wind gusts, tornadoes, and atmospheric electrical discharges (lightning).

Today it is more common to classify storms according to the characteristics of the storms themselves, and such characteristics depend largely on the meteorological environment in which the storms develop. Based on the thermodynamic, environmental characteristics, convective instability parameters, and physical processes, they are also classified as the air mass thunderstorms, squall lines, convective bands, mesoscale convective complex (MCC), convective clusters, mesoscale convective systems (MCS) (Markowski and Richardson 2010) (Fig. 19.2).

Air mass thunderstorms are also known as the local thunderstorms. They are mostly vertical in structure, are relatively short-lived, and usually do not produce violent weather at the ground. This type of storms is composed of one or more convective cells, each of which goes through a well-defined life cycle. Early in the development of a cell, the air motions are mostly upward, not as a steady, uniform stream but as one that is composed of a series of rising eddy. Cloud and precipitation particles form and grow as the cell grows. When the accumulated load of water and ice becomes excessive, a downdraft starts. The downward motion is enhanced when the cloud particles evaporate and cool the air – almost the reverse of the processes in an updraft. At maturity, the cell contains both updrafts and downdrafts in proximity. In its later stages, the downdraft spreads throughout the cell and diminishes in intensity as precipitation falls from the cloud. Isolated thunderstorms contain one or more convective cells in different stages of evolution.

Squall line thunderstorms (multicellular line storms) are many single cell storms arranged in a line. They are like mesoscale convective complexes in that they are a cluster of storms, but, instead of a circular-shaped cluster, they form in a multi-cell linear cluster. Squall lines form along cold fronts or as far as 80 miles ahead of a cold front. They form in a similar way to mesoscale convective complexes. If the air at or in front of the cold front is unstable, humid, and warm, squall line

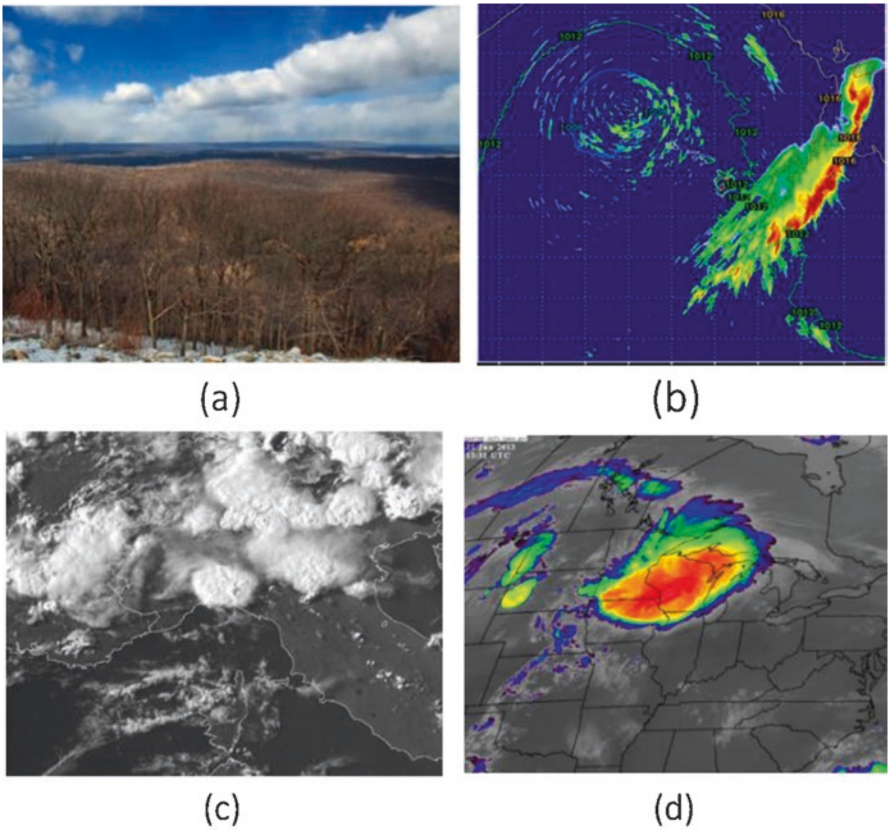

Fig. 19.2 Sequence of general appearance of (**a**) a squall line (Credit: Nicholas A. Tonelli from Northeast Pennsylvania, USA [CC BY 2.0 (https://creativecommons.org/licenses/by/2.0)]); (**b**) convective band; (**c**) mesoscale convective system (MCS); and (**d**) mesoscale convective complex (MCC)

thunderstorms can develop. They usually have heavy precipitation, strong winds, small hail, and lightning. In some rare cases, tornadoes can develop.

Mesoscale convective complexes are groups of thunderstorms that form in a large cluster that can span an entire state. Usually, they develop from a thunderstorm outbreak in the area of unstable air that is warm and humid. In the areas where they form, there is enough heat and moisture for air to rise and storms to form.

Multicell cluster are often associated with convective, vertical flow near the mountain slopes or to clearly express the cold fronts. Multicell clusters consist of a group of cells moving along. Most are short lived, but since these cells have formed in a cluster, the region in which they formed must have been conducive to thunderstorm formation. As such, individual cells will constantly form up and dissipate, but the cluster as a group will retain the same general characteristics. Multicell storms are usually more potent than single cell storms but are generally not severe.

Individual storms form in close vicinity of each other, and the storm cells feed off the outflow from other nearby storms. The result is a nearly circular, large cluster of storms that are often severe. They form in the late afternoon due to daytime heating and last into the night, but the most severe weather happens early in storm development. The storms can travel great distances before they dissipate and last at least 6 h.

Mesoscale convective system (MCS) is an organized system of thunderstorm, with excessive rainfall over larger mesoscale area and life cycle of several hours or more. They have a different form visible on satellite images. MCSs have regions of both convective and stratiform precipitation, and they develop mesoscale circulations in their mature stage. The upward motion takes the form of a deep-layer ascent as the result of latent heating and cooling in the convective region. A middle level layer of inflow enters the stratiform region of the MCS from a direction determined by the large-scale flow and descends in response to diabatic cooling at middle-to-low levels. A middle-level mesoscale convective vortex (MCV) develops in the stratiform region, prolongs the MCS, and may contribute to tropical cyclone development. The propagation of an MCS may have a discrete component but may further be influenced by waves and disturbances generated both in response to the MCS and external to the MCS.

19.5 Atmospheric Electricity

The phenomenon of atmospheric electricity in the clouds with vertical development is called atmospheric electricity or lightning. This is one of the oldest observed natural phenomena on Earth.

19.5.1 General Features of Lightning

Lightning is a transient, high-current electric discharge with path lengths measured in kilometres (Rakov and Uman 2003). It is a discharge of atmospheric electricity caused by formation of a strong electrical field in a thunderstorm cloud between the ice crystals, usually found in the cloud negative temperature region, near the cloud top, and negatively charged hail near the cloud base. The most common source of lightning is the electric charge separated in ordinary thunderstorm clouds or cumulonimbus clouds (see Cooray 2015). Well over half of all lightning discharges occur within the thunderstorm cloud and are called intracloud discharges. Scientists do not have complete knowledge as to how it occurs, or whether it is a reaction to sunlight affecting the upper atmosphere of the Earth or the electromagnetic field. Lightning can be seen in volcanic eruptions, extremely intensive in forest fires, surface nuclear detonation in heavy snowstorms in large hurricanes. It is often noticed during thunderstorms. Indeed, lightning and thunder, which manifests itself later, are an integral part of thunderstorms (Fig. 19.3).

Fig. 19.3 Atmospheric electricity, lightning

19.5.2 Origin of Atmospheric Currents

The source of the large negative currents must flow from the "top" to the surface of the Earth in order to keep charging it up negatively. Where is the main source for this negatively generated charge? Thunderstorms and lighting storms serve as battery which carries out negative charge to the Earth in large amounts. Thunderstorms charging the Earth with an average 1800 amperes, which is then being discharged through regions of fair weather. There are about 40,000 thunderstorms per day all over the world, and we can imagine them as batteries pumping the electricity to the upper layer and maintaining the voltage difference. However, thunderstorm estimates are very difficult to make because the lack of observations on the seas and oceans over all parts of the world to know the accurate number of thunderstorms. But according to some evidence, there are 100 lightning flashes per second, worldwide with a peak in the activity at 7:00 p.m. GMT.

19.5.3 The Mechanism of Charge Separation

We will discuss now about the most important aspect-development of the electrical charges. Based on many experiments including flying airplanes through thunderstorms, the charge distribution in a thunderstorm cell is something like that shown on Fig. 19.4. A typical feature of storms is the presence of turbulent area and the presence of strong updrafts and downdrafts currents (Cotton et al. 2010). Vertical upward currents transport small liquid water droplets from the lower regions of the storm to the higher layers of the cloud. Meanwhile, descendent currents transport hail and cloud ice from frozen upper regions of the storm.

When these collide, the water droplets freeze and release heat. This heat sometimes keeps the surface of things and ice slightly warmer than its surrounding environment and is taking up "wet hail," or "graupel." When this hail or groupel

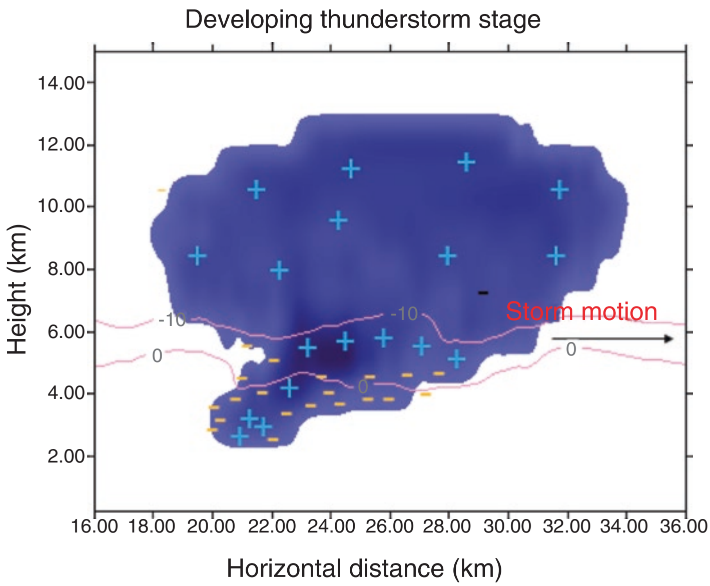

Fig. 19.4 Charge formation and the mechanism of charge separation in simulated storm

collides with additional water droplets and ice particles a critical phenomenon occurs. Electrons that deviate from ascendant particles are collected on descendant particles. Because electrons carry a negative charge, the result is a negative storm cloud electrification base and a positively charged top and the bottom a negative charge. Except for a small local region of positive charge in a bottom of cloud, which has caused everybody a lot of worry? One assumption is that it is probably due to a secondary effect of rain coming down. Anyway, the predominantly charge at the bottom and the positive charge at the top have the correct sign for the battery needed to drive Earth negative. The positive charges are 6–7 km upper in negative temperatures about −20 °C, whereas the negative charges are 3 or 4 km high in temperature ranges between 0 and − 10 °C. The charge at the bottom of the cloud is large enough to produce potential difference of 20–100 million volts between the cloud and the Earth and thus much bigger than the 0.4 million volts from the "sky" to the ground in a clear atmosphere. These large voltages break down the air and create giant arc discharge. When the breakdown occurs the negative charges at the bottom of the thunderstorm are carried down to the Earth in the *lightning strokes*. Now we will describe in some detail the character of lightning. There is large voltage difference around (one piece of cloud and another, or between only cloud and another or between the cloud and ground), so that the air brakes down. In each of the independent discharge flashes, the lightning strikes that we are seeing there are approximately 20 or 30 coulombs of charge brought down. It takes a thunderstorm only 5 sec after each lightning stroke to build its charge up again. Another important aspect is that under certain conditions electric field can have considerable influence

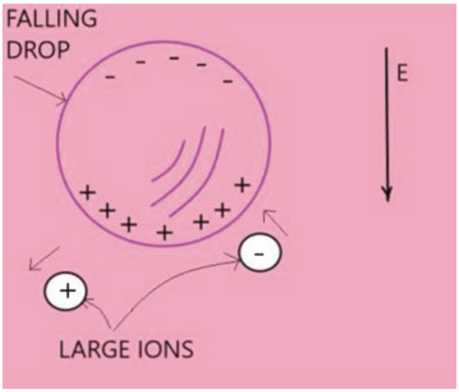

Fig. 19.5 Formation of charge of falling drop

on the drops. The strokes occur irregularly, but the important point is that it takes about 5 seconds to recreate the original condition. Thus, there are approximately 4 amperes of current in the generating machine of the thunderstorm. There are two theories which have been invented to account for the separation of the charges in a thunderstorm: Breaking-drop theory and Wilson theory. All the theories involve the idea that there is some charge on the precipitation particles and different charge in the air. Then by the movement of the precipitation particles, the water or the ice through the air there is separation of electric charge (Fig. 19.5).

19.5.4 Creation of a Field

The most common source of lightning is the electric charge separated in ordinary thunderstorm clouds (cumulonimbus). The atmosphere is a very good insulator that prevents the electrical flow, so a huge amount of charge should be created before lightning can appear. When the threshold has reached the charge, the strength of the electric field exceeds insulating properties of the atmosphere and lightning appears. Electric field inside the storm is not unique, and it is developed there. Under negative discharge of storm base, the surface of the Earth begins to form positive charge. This positive charge will follow the storm wherever it moves, and it is responsible for lightning from cloud to Earth. However, the electric field inside the storm is much stronger than the one between the base of the cloud and the Earth's surface, so lightning mostly (~75–80%) occurs within the storm. The storm which moves accumulates another region with positively charged particles along the ground. As a result of differences in charges, positively charged particles rise high objects such as trees, houses, and telephone lines. Those channelled with a negative charge will be lowered from the bottom of the storm to the Earth's surface in a series of rapid impulses, invisible to the human eye. As a leading negative charge is approaching, positive charge collects at the ground and the objects found there. This positive charge goes to the negative charge and "lightning" forms. When these channels are merged, electric transmission appears as a result, which is visualized as atmospheric lightning.

19.5.5 Lightning Stroke

Lightning stroke is the direct discharge of an electrical charge between the atmosphere and the object of Earth. It is a sudden flow of electric charge between the electrical charge area of a cloud also called intra-cloud and another cloud called (CC lightning) or between the charged cloud and the ground (CG lightning). Lightning is caused by electricity moving in between clouds or between the clouds and the ground.

19.5.6 Mechanism of the Lightning Discharge

The process of ionization and polarization are dominant in the formation of charged suspended particles and water drops in the air which also causes charging of clouds (Ćurić 2001b). There is a difference in distribution of electric charge in a calm sky conditions and clouds. While in a calm sky, the positive (+) and negative (−) charges are evenly spaced throughout the atmosphere causing a neutral charge, in thunderstorm due to a presence of different type of hydrometeors.

The ice crystals have a positive charge, and the hailstones have a negative charge. An updraft moves the ice crystals to the top of the thunderstorm cloud, while the hailstones are moved to the bottom of the thunderstorm by its downdraft (Fig. 19.6).

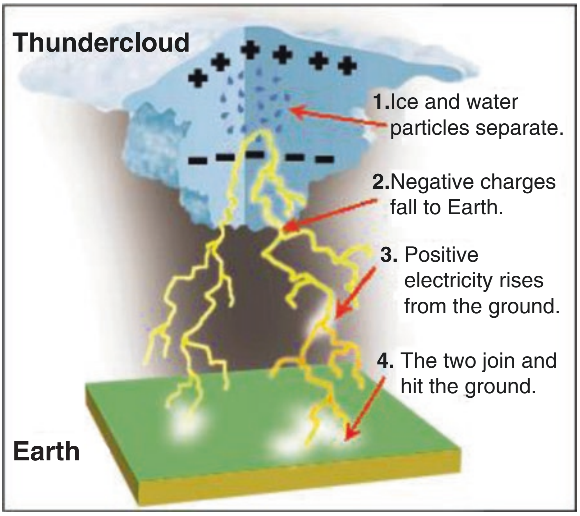

Fig. 19.6 Lightning formation. Credit: Pearce, (1989)

These processes initiate the charge separation into two levels: the positive charge at the top and the negative charge at the cloud base that induces a positive charge at the ground level. As the result of the potential difference in the cloud the initial discharge originates in the cloud called *pilot leader*. The first discharge moves to earth in steps of about 50 metres, therefore, termed the *stepped leader*. When the pilot leader reaches near the Earth, the electric field intensity increases, and due to this, the charges of an opposite polarity in the form of a short steam rises from the earth to meet the tip of downward leader. When a contact is made between the pilot leader and the short upward steamer, a return streamer travel from the Earth to cloud along the ionised channel formed by the pilot leader. When the stepped leader and the positive charge from the Earth meet, a strong electric current carries positive charge up into the cloud. This electric current is known as the return stroke that moves very fast producing well known, intensely luminous *lightning flash*.

19.6 Tornado

Cyclones can produce tornadoes that complement destructive power of storm (Fig. 19.7). Tornadoes are often embedded in rainy zones in the right frontal quadrant, well away from the centre of tropical storm in terms of movement. A thunderous tornado is a stormy wind in the form of rotating vortex which extends downwards from a cumulonimbus cloud.

Tornado is a very prominent rotating column of air (usually counterclockwise in the Northern Hemisphere), which descends from thunderstorm cloud, and encounters Earth. Although tornadoes are usually short in duration, only a few minutes, they can sometimes be manifested more than an hour and travel several miles

Fig. 19.7 Tornado over land

causing considerable damage. Tornado consists of drops of water, sand, and dust from various objects which are raised from the soil with the help of strong wind. Tornadoes often occur in the United States and Mexico and usually move with speed of up to 50 km/hour. Vortex in the air rotates at a speed of 160 km/hour. If a tornado occurs at sea, it is called waterspout or thrombus and lasts from several minutes to an hour. Tornadoes inflict heavy damage and leave a wasteland behind. The United States experience more tornadoes than any other country. During the tropical year around 1000 tornadoes affect the United States. The peak of the tornado season is April–June. They are characterized by continuous vertical rotation and formation of strong vertical shearing in the surrounding air. Searing represents a change of wind speed and/or direction with height. Vertical movement raises rotating column of air generated by shear of wind speed. This provides two different rotations in supercell storm, cyclonal rotation, or counterclockwise rotation and anticyclonic rotation or clockwise rotation. Oriented wind shear intensifies cyclonic rotation and reduces anticyclonic rotation (rotation to the right side of the vertical movement). Supercell storm is a rotating thunderstorm. The most dangerous products of thunderstorms are tornadoes. By definition, a tornado is a strongly rotating column of air that spreads from the storm towards the ground. A tornado is characterized by:

- Winds speeds of 64–176 km/hour
- Radius, which changes from a few hundred metres up
- Wind speeds greater than 480 km/h
- Counterclockwise rotation in the Northern Hemisphere and vice versa in the Southern Hemisphere

19.6.1 *Classification of Tornadoes*

The most common acceptable scale for determining the damage is the Fuji's scale (F) for classification of tornadoes, which is based on the damage that the tornado has caused. A recently revised scale is used to assign the speeds of the tornado in 1971, known as enhanced Fuji's (EF) (Table 19.1). Based on their physical characteristics, there are several different basic types of tornadoes: multilevel rotating tornado, accompanying tornado and tornado over water.

Table 19.1 Classification of tornadoes according to EF scale

EF scale	Class	Wind speed (km/h)	Description
EF0	Weak	105–137	Storm
EF1	Weak	138–177	Moderate
EF2	Cool	178–217	Significant
EF3	Cool	218–266	Dangerous
EF4	Strong	267–322	Destructive
EF5	Strong	> 322	Catastrophic

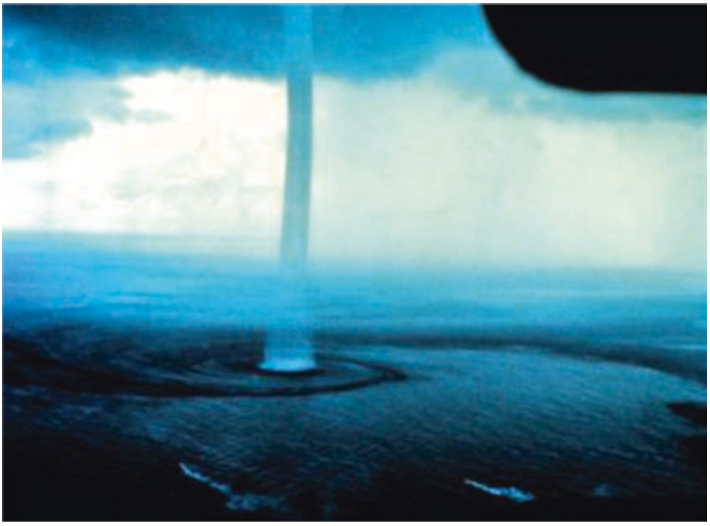

Fig. 19.8 Tornado over water

Multilevel rotating tornado: a tornado in which there are two or more columns of vortex air, which rotates around a common centre. This tornado is often controlled with intensive tornadoes. These spins often form small areas of damage expressed along the main path of the tornado.

Accompanying tornado: a weak tornado formed very close to a large, strong tornado. It is possible that the accompanying tornado rotates around the strong tornado.

Waterspout: tornado that occurs over water surface (Fig. 19.8). This type of tornado can be divided into two smaller types: a fountain of fine weather and a fountain in the tornado.

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Chapter 20
Meteorological Hazards

20.1 Natural Disasters

Natural hazards are severe and extreme weather and climate events. Although they occur in all parts of the world, some regions are more vulnerable to certain hazards than others. Natural hazards become disasters when people's lives and livelihoods are destroyed. According to the main driving force of the event, natural hazards are classified into four main groups: geological hazards, meteorological hazards, hydrological hazards, and biological hazards.

1. Geological hazards occur because of geological processes such as movement in the tectonic plates and volcanic activity. These events include earthquakes, volcanic eruptions, and landslides.
2. Meteorological hazards occur as a result of processes in the atmosphere (Ahrens and Samson 2011). Meteorological hazards include extreme temperatures, heat waves, cold spells, hurricanes, tornadoes, droughts, and severe storms.
3. Hydrological hazards are hazards involving water processes. Examples include floods, droughts, and tsunamis.
4. Biological hazards occur due to the biological processes of the earth and primarily involve the spread of diseases and pests. Epidemics, pandemics, and airborne diseases all fall into the biological hazard's category.

Each year, natural disasters cause great material damage in various areas of the Earth's Planet. Figure 20.1 shows the distribution of natural hazards, based on statistical evidence for the period 1980–2005. According to data provided from the World Meteorological Organization (WMO), the following percentage of hazards was estimated: floods (32%), storms (22%), epidemics, hunger (17%), dry (9%), earthquakes (9%), landslides (5.3%), extreme temperatures (2.5), fire (1.9%), volcanoes (1.8%), and tsunami (0.5%). It is obvious that nearly 90% of all-natural disasters are from meteorological origin related to weather, climate, and water. Human and material losses caused by such disasters are a major obstacle to

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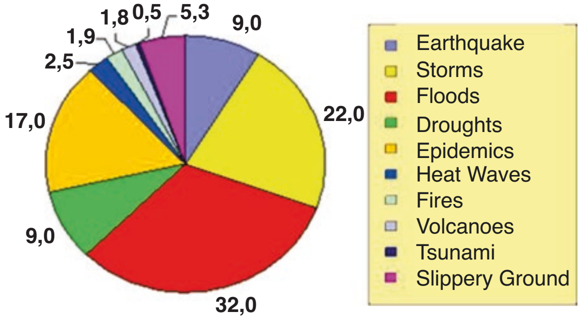
Fig. 20.1 Distribution of the natural hazards, according to the statistical data evidence 1980–2005



sustainable development. By issuing accurate forecasts and warnings in a form that is readily understood and by educating people on how to prepare against such hazards, before they become disasters, lives and property can be protected. Emphasis is on disaster risk reduction: one dollar invested in disaster preparedness can prevent seven dollars' worth of disaster-related economic losses – a considerable return on investment. Having regard to the amount of damages and their continuity on the Earth as a whole, the United Nations General Assembly designated the 1990s as the International Decade for Natural Disaster Reduction (IDNDR). Its basic objective was to decrease the loss of life, property destruction, and social and economic disruption caused by natural disasters, such as earthquakes, tsunamis, floods, landslides, volcanic eruptions, droughts, locust infestations, and other disasters of natural origin. The decade was intended to reduce, through concerted international action, especially in developing countries, loss of life, property damage, and social and economic disruption caused by natural disasters. The first objective was to emphasize the significance of these phenomena for human life and properties. It was recommended for each country, the experts to make efforts and to address this problem associate with the natural disasters by individual study of a given area or a region. From the obtained knowledge they need to develop and formulate procedures for continuous monitoring of the evolution of a given phenomenon. The next step is determination of the thresholds of various categories of natural disasters and definition of procedures for early warnings. The goal of the above-mentioned procedures is to reduce the number of fatalities and to mitigate the material damages by natural disasters. In this systematic process of disasters examination, two important elements were considered significant. First, some natural disasters can be only monitored and announced. Second, some disasters can be monitored in order to take appropriate actions and reduce their harmful effect. In the first case, it is a passive measure of protection, while in the latter case, an active measure of protection. According to the recommendation of the United Nations, many countries have taken serious steps towards meeting the goal of reducing damage from natural disaster. For example, to specify that Mexico, Canada, and the United States formed a team of 50 scientists in order to formulate a project for reducing damage from elemental disasters in 3 years. Recently, as signatories to the Sendai Framework for Disaster Risk Reduction 2015–2030, WMOs Members have undertaken to prevent new and reduce existing disaster risk through the implementation of a range of integrated and inclusive measures that prevent and reduce hazard exposure and vulnerability to disaster, increase preparedness for response and recovery and thus strengthen resilience.

20.2 Meteorological Hazards

Disasters caused by extreme weather conditions are referred to as meteorological disasters (Spiridonov and Ćurić 2011; Spiridonov et al. 2013). Such disasters are usually related to sudden and adverse changes in the weather or weather-forming processes. Extreme heat, excessive rainfall, and strong winds affecting the Earth's atmosphere in a negative manner and causing negative effects are called as meteorological disasters.

20.2.1 Droughts

Droughts are a prolonged dry period in the natural climate cycles that can occur anywhere in the world. It is a slow onset phenomenon caused by a lack of rainfall. When drought causes water and food shortages, there can be many impacts on the health of the population, which may increase morbidity and result in death. The main reason for any drought is a deficit of rainfall or extended periods without the occurrence of rainfall (Fig. 20.2). This deficit results in water shortage for some activity, some groups or some environmental sectors. Drought is also related to the period of rainfall deficit in some specific area. Other climatic factors such as high temperature, strong wind, and low relative humidity are often linked to drought. Drought differs from other disasters that develop slowly, sometimes for months and years. Drought is a characteristic of the climate feedback. It occurs in virtually all climatic zones, and its characteristics change significantly between regions. Drought differs from aridity in that it is an occasional feature, and aridity is a constant characteristic of regions with little rainfall. Drought is more of a natural phenomenon, or more of a natural disaster. Experience of droughts confirms the sensitivity of human communities to this natural disaster. There are two kinds of drought definitions:

1. Conceptual definition
2. Technical definition

Conceptual definitions help to understand the importance of drought and its effects and damage to agriculture, food production, and others. Operational definitions help to identify the beginning and end of droughts and the degree of danger. They also specify the degree of deviation from average rainfall in a climate period. There are different types of drought, which are briefly described hereafter:

Fig. 20.2 Drought (Credit: Tomas Castelazo, www.tomascastelazo.com / Wikimedia Commons / CC BY-SA 4.0)

Meteorological drought is defined on the basis of the degree of aridity, compared to the average normal value, and based on the duration of the dry period. Meteorological definitions of drought must be region specific, because the atmospheric conditions that affect the lack of rainfall are largely specific to each region.

Agricultural drought links various characteristics of the meteorological drought to agricultural impacts, focusing on the lack of rainfall, the difference between actual and potential evapotranspiration, the deficit of water in the land, the reduced level of groundwater and or surface water levels in reservoirs, and others.

Hydrological drought is associated with continuous low flow and/or volume of water in catchments and reservoirs, which lasts for months or years. The hydrological drought is a natural phenomenon that can be worsened by human activity. Hydrological droughts are usually linked to meteorological droughts and their recurrence interval of compliance changes. Changes in land use and degradation of the same may affect the size and frequency of hydrological drought.

Socio-economic drought links the provision and the need for some economic goods with elements of meteorological and hydrological and agricultural drought. It differs from other types of drought that its occurrence depends on the process of assistance and requests. Energy supply with many economic goods, such as water, food, grains, fish, and hydropower energy, depends on weather. As a result of natural variability of climate, water supply is extensive in some years but not enough to satisfy human and environmental needs in other years. Socio-economic drought occurs when the demand for economic goods exceeds supply as a result of the collapse in the water supply associated with the weather.

20.2.2 Tropical Cyclones

A tropical cyclone represents a system with low air pressure in the centre and organized circulation that develops over tropical and sub-tropical ocean waters. Its horizontal dimension is usually 200–2000 km. A tropical cyclone is characterized by a warm core in the centre, very steep pressure gradient, and strong cyclonic (clockwise direction on the Southern Hemisphere), and very strong destructive surface winds, with a maximum speed of (300 km/h). Large storm waves rise with the wind causing severe flooding in coastal areas. That is especially occurs when the storm surge coincides with normal high waves. Hurricanes are known for their destructive winds. Cyclonic wind causes great damage to marine traffic and infrastructure on the mainland. Many homes were damaged and destroyed when strong wind simply raised the roofs of houses. Tropical cyclones can also cause flash flooding, urban, or river flooding. For example, Fig. 20.3 shows a numerical simulation of an absolute vortex field at 700 mb, of the tropical storm Pabuk on 5 Jan 2019 03UTC, which hit coastal villages and world-famous tourist resorts on southern Thailand's east coast causing flooding, winds and surging seawater.

20.2.3 Floods and Flash Floods

Weather systems such as cyclones, storms, fronts, or other forms of clouds can, under certain unstable atmospheric conditions, increase their frequency and strength and cause intense rainfall and flooding. Heavy rainfall is the most frequent adverse weather phenomena. Floods occur in the case of intense torrential rainfall, when there is significant accumulated rainfall or extremely high daily rainfall in a few hours (Fig. 20.4).

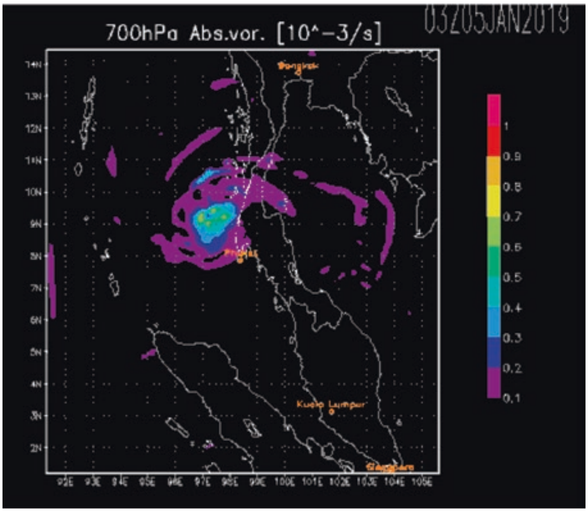


Fig. 20.3 Numerical simulation of tropical storm (WRF-ARW forecast of tropical storm Pabuk that hit Pucket on 3 Jan 2019)

Fig. 20.4 Skopje flash flood on August 6, 2016. (Credit: Ministry of Interior, Macedonia)

A single extreme weather event cannot be attributed to climate change. However, according to the analysis of statistical trends, climate change is expected to increase the intensity of the water cycle in many parts of the world, causing more droughts and more floods. Floods occur in all possible forms, from small sudden floods to waterspouts that cover large areas. Floods result from the interaction of precipitation, surface runoff, snow melting, evaporation, wind, and local topography. In inland areas, the flood regime varies greatly depending on the size of the basin, topography, and climatic characteristics. One of the most common incidences of flash flooding occurs when there is a heavy period of rainfall following a period of drought. This is because the ground is very dry and cannot absorb so much water at once. Heavy rainfall can lead to several risks, to human life, damage to buildings and infrastructure, and loss of crops and livestock. Landslides that could threat human lives disrupt transport and communications and cause damage to buildings and infrastructure. Extreme storms bring heavy rainfall or hail fall, severe winds, and occasional snowfall. Thunderstorms cause sudden electrical discharges in the form of lightning and thunder. In some parts of the world, they cause tornadoes. Violent storms and flash floods, besides the direct consequences in terms of human health, have indirect effects, such as increased stomach and intestinal illness among evacuees, and an impact on mental health, depression, and post-traumatic disorders. The consequences of the damages from extreme storms and floods sometimes include fatalities and injuries, but these usually are counts of drownings, vehicle accidents, and other immediate injuries. Such extremely severe weather developed on August 6, 2016, and affected a wider urban and rural areas of Skopje. The supercell storm (Spiridonov and Ćurić 2019) was associated with torrential rainfall, wind gusts, strong lightning, and flash flooding. During this disastrous heavy rainfall event, 23 people lost their lives with huge material and economy damage and injuries. Less reported, but often more numerous, are illnesses and even deaths from consumption of contaminated water. Neglecting direct and indirect health impacts may result in inadequate preparation for an appropriate response in future extreme weather events. Embankments (dams) may flood because they cannot withstand the blows of rising water in rivers and streams that carry large amounts of melted snow.

Floods caused by heavy storms threaten the lives of people and goods around the world. Floods contaminate sources of fresh drinking water, increasing the risk of diseases transmitted through water, and create conditions for the propagation of diseases that are transmitted by insects such as mosquitoes. They also cause drowning and physical injury, damage to homes, and disrupt the supply of medical and health services.

20.2.4 Landslides or Slippery Ground

Landslides or slippery ground represent a gravity-initiated process of transport of mainly dry material (e.g. rock, debris, sand, soil, or sediment) downslope. It is considered as local scale phenomena which occur suddenly and cause severe threats and damages. This type of hazard does not explicitly belong to the group of meteorological disasters, but certain weather events associated with heavy rainfall, flash-flooding, and rapid snow melting with combination of warming and rainfall can accelerate the process and can serve as a triggering factor of landslide. Landslide complexes, specific mountain slopes, river valleys, and dams are as most vulnerable to landslides and favourable conditions for the generation of debris flows and flash-floods.

A landslide happens as the result of a disbalance between gravity force and friction which causes movement of the material downhill. Precipitation significantly fastening the process due to larger weight to the slope and consequently increase of gravitational force making it easier to move material downhill. These processes are more dominant during the rainy season and especially common during severe storms.

20.2.5 Avalanches

Avalanche is a mass of snow and ice that suddenly falls down the mountain slope, often carrying side, rocks, and waste (Fig. 20.5). Avalanches can be very destructive, with speeds that exceed (150 km/h). The snow that is moving also pushes air forward so that the avalanche wind becomes strong enough to cause major damage to infrastructure, buildings, forests, and mountain resorts. Thousands of avalanches occur each year, and they kill an average of 500 people worldwide.

20.2.6 Desert and Dust Storms

Desert and dust storms are a set of dust or sand particles, which can be collected to major heights by strong and turbulent wind (Fig. 20.6). They occur mainly in parts of Africa, Australia, China, and the United States. They affect the lives and health of people, especially busy people in the open away from shelter.

Fig. 20.5 A powder snow avalanche in the Himalayas near Mount Everest

Fig. 20.6 Dust storm. (Credit: Roxy Lopez [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)])

They have a particularly negative impact on transport, because visibility is reduced to just a few meters.

20.2.7 Heat Waves and Cold Spells

Heat waves represent the most dangerous phenomena in regions of high latitudes, where there are extreme values of temperature and humidity over a longer period which may last several days in the warmest months. A favourable condition for the occurrence of heat waves are anticyclones, calm, and fair weather, expressed a high air pressure and advection of hot air and the occurrence of very high temperature. Stable, slowly moving and stagnant-air mass in the urban environment can cause deaths, particularly of the youngest, and the older and infirm people. In 2003 much of Western Europe was under the influence of heat waves during the summer months. In France, Italy, the Netherlands, Portugal, Spain, and Britain, they have caused about 40,000 losses of human lives. Extreme cold waves cause hypothermia and circulatory and respiratory diseases.

20.2.8 Severe Storms, Lightning, and Tornadoes

Dangerous storms are a source of sudden atmospheric electrical discharges that are manifested in the form of lightning and thunders (Fig. 20.7). They often bear strong heavy rainfall or strong winds and occasional snow. In some parts of the world, tornadoes occur regularly. Tornadoes are especially common in the great plains of the North American continent, but they also appear in temperate latitudes. They can cause enormous damage. Flash floods are accompanying phenomena of tornadoes. Widespread lightning during dry periods is an important factor in the initiation of wildfires in forests and grasslands.

20.2.9 Forest of Wildland Fires

Massive and devastating fires can be initiated by lightning or by human activity during and after periods of drought in almost all parts of the world. Favourable conditions for the formation of fire (Fig. 20.8) are very hot and dry, high temperatures, and low relative humidity. Occurrence of wind in the atmospheric conditions could further inflame and reinforce the existing fire. When fires destroy forests, pastures, and corn, they kill livestock and wild animals, and they damage or destroy the settlements and put at risk the lives of the residents.

Fig. 20.7 Dust storm (Credit: Roxy Lopez [CC BY-SA 3.0 (https://creativecommons.org/licenses/by-sa/3.0)])

Fig. 20.8 Forest fires (wildfires)

Fig. 20.9 Heavy snowfall

20.2.10 Heavy Rainfall and Snowfall and Strong Winds

Heavy rain and snow weather are dangerous for vulnerable groups (Fig. 20.9). They cause problems in the traffic, the infrastructure, and the communication networks.

The accumulation of snow can cause destruction of roofs at the buildings. Strong winds are hazardous to aviation, sailors, and fishermen, also for high buildings such as towers, cranes, and skyscrapers. Blizzard conditions are strong storms accompanied by low negative temperature, with strong winds and snowing. They are dangerous to humans and livestock. Blizzard conditions cause airports to be closed for taking off and landing operations and cause chaos in the streets and sidewalks.

20.2.11 Extreme Air Pollution

Atmospheric pollutants include matters and harmful gases from industry, from vehicles and human activities. Smoke and haze arise as a result of forest fires, cleansing grains, or ash from volcanic explosions in stable conditions. Smoke, haze, and

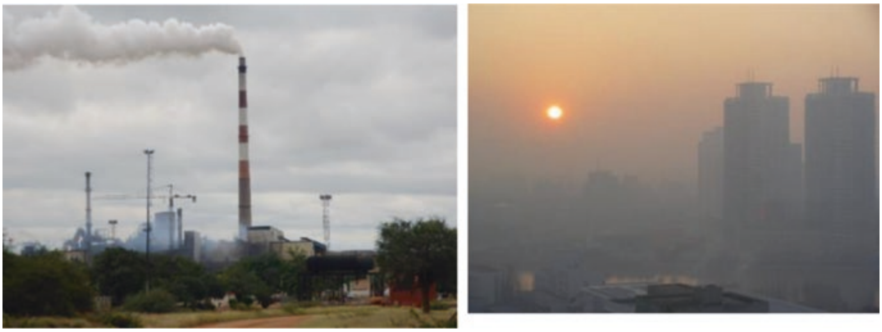


Fig. 20.10 Air pollution in urban areas and industrial zones

pollution have a serious impact on human health, especially in urban and industrial areas (Fig. 20.10). They reduce visibility and thus air and road traffic may be interrupted. Smoke, acid rain, ground floor harmful ozone, the ozone hole, and the adverse growth in the greenhouse effect are also caused by air pollution. A persistent high-pressure system associated with stable atmospheric conditions and temperature inversion often leads to pollution accumulation and appearance of extreme air pollution episodes with high concentration, poor air quality and harmful health impact. That is usually case in the urban areas with a specific topography, environmental conditions, urban structure, and uncontrolled emissions.

20.2.12 Tsunami

Tsunami is a wave phenomenon generated by a meteorological or atmospheric disturbance. The term “tsunami” is taken from the Japanese tsunami, which means “harbour wave”. They are initiated by moving air pressure disturbances associated with atmospheric gravity waves, pressure oscillation, frontal passages, squalls, gales, typhoons, hurricanes, and other atmospheric sources. These disturbances generate barotropic ocean waves in the open ocean through resonance, which can subsequently be amplified near the coast through shoaling and basin resonance. In general tsunami is a series of ocean waves, which are created as a result of underwater earthquakes, landslide, or volcanic eruptions. Although tsunami belongs in the category of a marine geo-hazard, such severe ocean waves have devastating for the coasts and their communities. They can also have lasting and damaging effects on the coastal landscape, causing long-term coastal erosion, and on marine ecosystems. These waves may reach large sizes and have enough energy to move oceans across and manifest negatively when they reach the mainland. Tsunami is common in the Indian Ocean (Indonesia, Sumatra, Japan, etc.). Because of the enormous water mass and energy, which they carry with them, tsunami waves can destroy whole coastal regions. The number of victims can be very high, because it creates

Fig. 20.11 Tsunami (tidal ocean waves)

huge waves which are flooding the coastal areas and destroying infrastructure (Fig. 20.11).

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Chapter 21
Atmospheric Optics

Before to start with description on the atmospheric optics, we will briefly explain the basic principles of energy which comes from the Sun and the radiation. The energy emitted by the Sun provides almost (99.9%) of the total energy that heats the Earth's surface. The large changes in solar heat run the winds in the atmosphere and currents in seas and oceans. Therefore, first it is necessary to understand the way the Sun heats the Earth, and this warming is changing the geographical space and time. For this purpose, we primarily study the heat and the temperature as variables, and the mechanisms for their transfer.

21.1 Basic Properties of Sunlight

Sunlight is characterised by very high intensity and continuous movement. The brightness of direct sunlight varies by season, time of day, location, and sky conditions (Zuev 1995; Andrews 2019). The basic properties of light when it passes through different environments in the atmosphere are:

- Reflection
- Refraction
- Scattering
- Diffraction
- Interference

Skylight is characterised by sunlight scattered by the atmosphere and clouds, resulting in soft, diffuse light. In a cloudy climate, the diffuse sky is often the main source of useful *daylight*. If a single wavelength is present, we say that we have *monochromatic light*. If all wavelengths of visible light are present, our eyes interpret this as *white light*. If no wavelengths in the visible range are present, we interpret this as *dark*.

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21.1.1 Light Velocity and Refractive Index

The energy of light is related to its frequency and velocity as follows:

$$E = h\nu = hC/\lambda \tag{21.1}$$

where E is energy, h - Planck's constant, 6.62517 x 10**-27 erg.sec, ν – frequency, C is a speed of light which is equal to 2.99793 x 10**10 cm/sec, and λ is a wavelength. The speed of light, C in a vacuum is 2.99793 x 10**10 cm/sec. Light cannot travel faster than this, but if it travels through a substance, its velocity will decrease. Note that from the equation given above:

$$C = \nu\lambda \tag{21.2}$$

The frequency of vibration, ν, remains constant when the light passes through a substance. Thus, if the velocity, C is reduced on passage through a substance, the wavelength, λ, must also decrease. We here define refractive index, n, of a material or substance as the ratio of the speed of light in a vacuum, C to the speed of light in a material through which it passes, C_m.

$$n = C/C_m \tag{21.3}$$

Note that the value of refractive index will always be greater than 1.0, since Cm can never be greater than C. In general, Cm depends on the density of the material, with Cm decreasing with increasing density. Thus, higher-density materials will have higher refractive indices. The refractive index of any material depends on the wavelength of light because different wavelengths are interfered with to different extents by the atoms that make up the material. In general, refractive index varies linearly with wavelength.

21.1.2 Dispersion

The velocity of light in a material, and hence the index of refraction of the material, depends on the wavelength of the light. Since the refractive index depends on the wavelength of the light, light waves with different wavelengths and therefore different colours are refracted through different angles. This is called dispersion, because white light is dispersed into its component colours while traveling through the material. Snell's law combined with a wavelength-dependent index of refraction **n** explains the dispersive properties of a prism (Fig. 21.1).

The sides of a prism are not parallel and light changes direction when it passes through it. A ~ 1% variation in the index of refraction over the entire visible range of electromagnetic radiation still results in a significant change in the direction of the emerging red and blue rays. Since in general the index of refraction is bigger for

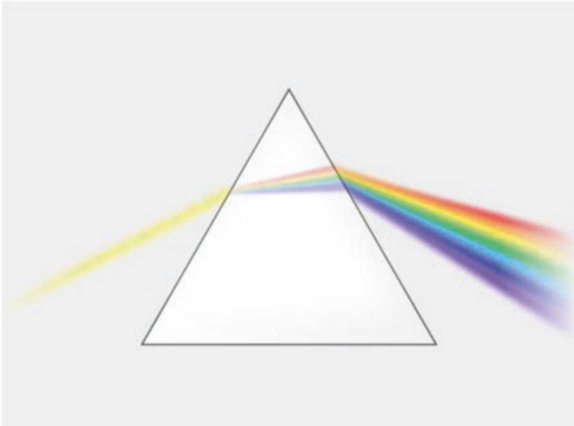


Fig. 21.1 Light dispersion mechanism

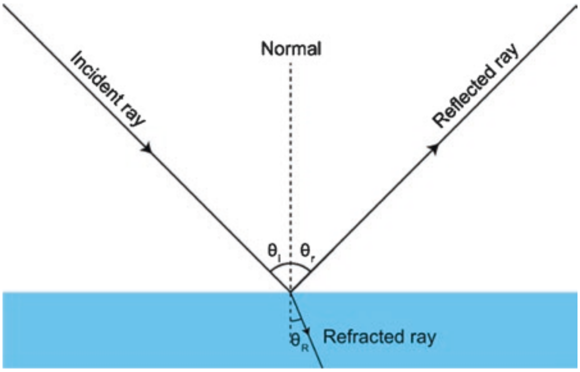


Fig. 21.2 Reflection

shorter wavelengths, blue light bends more than red light. Starlight passes from vacuum where refraction index n = 1.000, while into the atmosphere, n = 1.0003. As density of air increases at lower altitudes n changes. As n increases, light is bent continuously, deflecting light. From all colour a blue light bent the most.

21.1.3 Reflection

The law of reflection states that when light rays are reflected, they always bounce off the reflecting surface at the same angle (the angle of reflection) at which they meet that surface (the angle of incidence). Internal reflection occurs when light that is traveling

through a transparent material, such as water, reaches the opposite surface and reflects into the transparent material. Internal reflection is an important factor in the formation of optical phenomena, such as rainbows. Figure 21.2 shows a graphical illustration of the reflection process.

21.1.4 Refraction

Refraction is the bending (reflecting) of light due to a change in velocity as it passes obliquely from one transparent medium to another (see Fig. 21.3). Furthermore, light will also gradually bend as it traverses a material of varying density. The bending of light by refraction is responsible for such common optical illusions as the apparent displacement of the position of the stars, Moon, and Sun.

21.1.5 Scattering

In the air, part of the sunlight is scattered. The small particles (molecules, tiny water droplets and dust particles) scatter photons the more, the shorter their wavelength is. Therefore, in the scattered light, the short wavelengths predominate, the sky appears blue, while direct sunlight is somewhat yellowish, or even reddish when the sun is very low. Light is scattered, generally in all directions when it passes aerosol particles or air molecules. Scattering of sunlight in the atmosphere (Fig. 21.4) can be categorized into two types: Rayleigh and Mie scattering. As it is described previously in Chap. 5, *Rayleigh scattering* refers to the light scattering from the molecules of the air and can be extended to scattering from particles up to about a tenth of the wavelength of the light, resulting in appearance of a blue sky. *Mie scattering* is caused by pollen, dust, smoke, water droplets, and other particles in the lower portion of the atmosphere. It occurs when the particles causing the scattering are larger than the wavelengths of radiation.

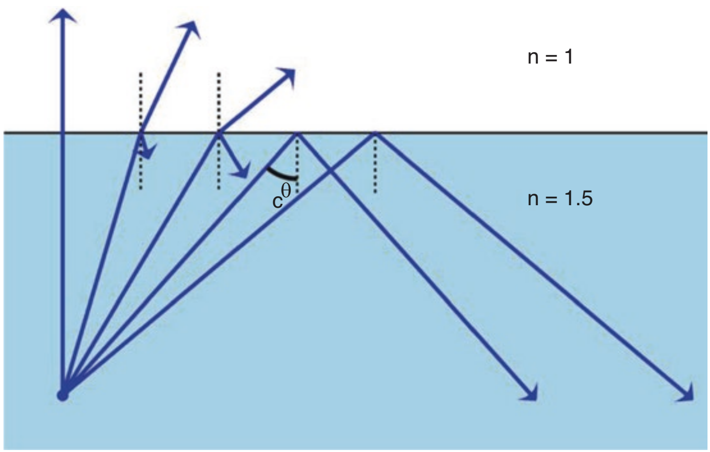
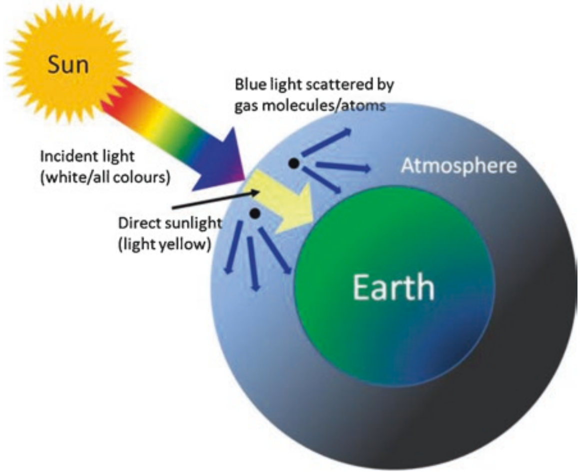


Fig. 21.3 Refraction

Fig. 21.4 Schematic illustration of scattering

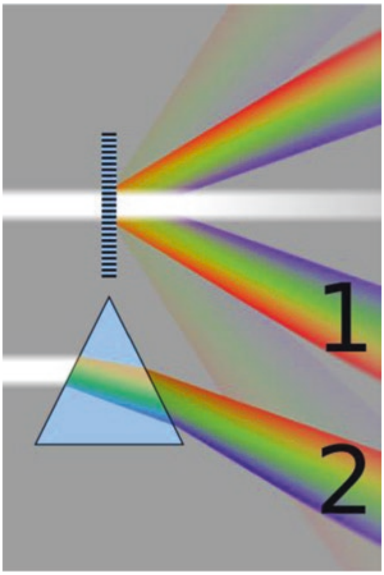


21.1.6 Diffraction

Diffraction is a physics concept. It occurs when waves bend around small obstacles, or when waves spread out after they pass through small openings. Generally, diffraction effects are most pronounced when the dimensions of the obstacle nearly agree with the wavelength of the wave. Diffraction occurs with all waves, including sound waves, water waves, and electromagnetic waves such as light that the eye can see. Augustin-Jean Fresnel (1788–1827) was a French engineer and physicist who contributed significantly to the establishment of the theory of wave optics. Fresnel studied the behaviour of light both theoretically and experimentally. There are some differences between diffraction and refraction. As it is shown on Fig. 21.5, diffraction is bending or spreading of waves around an obstacle, while refraction is bending of waves due to change of speed. Both are wavelength dependent.

However, diffraction produces a colorize pattern, whereas refraction creates visual illusions but not fringe patterns and can make objects appear closer than they really are, but diffraction cannot do that. Light interference. In physics, interference is a phenomenon in which two waves superpose to form a resultant wave of greater, lower, or the same amplitude. Interference effects can be observed with all types of waves, for example, light, radio, acoustic, surface water waves, or matter waves. In physics, absorption of electromagnetic radiation is the way in which the energy of a photon is taken up by matter, typically the electrons of an atom. Thus, the electromagnetic energy is transformed into internal energy of the absorber, for example, thermal energy. When a light wave with a single frequency strikes an object, several processes occur. The light wave could be absorbed by the object, in which case its energy is converted to heat.

Fig. 21.5 Diffraction



21.2 Photometeors

Photometeors are atmospheric optical phenomena that occur as the result of physical processes (e.g. reflection, diffraction, scattering, interference) and their interaction with sunlight or moonlight under some particular circumstances (Timofeyev and Vasilev; Ćurić and Janc 2016; Spiridonov 2010; Spiridonov and Ćurić 2011; Heidorn and Whitelaw 2010). The most common examples include halos, rainbows, mirages, cloud iridescences, glories, coronas, sun dogs, and aurora.

21.2.1 Mirage

Mirage is an optical effect caused by atmospheric refraction of light when it passes from the medium (air) of a certain density to the medium (air) of another density. In this process, the object appears displaced from its true position (Fig. 21.6). Light travels through air of different densities at different speeds. The density of air changes with temperature, with cold air being denser. Near a surface such as a hot road, or an ice field, there can be a strong temperature gradient which produces a density gradient which causes the air to act like a lens near the surface. However, these two examples are very different. In the case of the ice field, the density gradient will cause surface features a long distance off to appear stretched upwards. While in the case of the hot road the density gradient will be in the other direction which results in the sky being reflected, or bent back up, towards the observer just near the hot surface. The result can be the false appearance of water in the distance

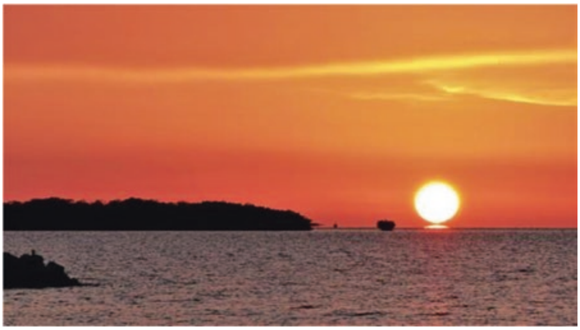


Fig. 21.6 Mirage

and thus the most commonly known idea of a mirage. The mirage known as low mirage occurs when the image appears below the actual location of the observed object. During the phenomenon of vision, objects sometimes appear as if floating above the horizon. Appearance is considered as the upper mirage as the image seen above its real position. Mirage, which changes the visible side of the object, is called ascent. A type of ascent, known as fatamorgana is often observed in coastal areas, lofty towers that appear out of thin air layer.

21.2.2 Rainbow

The most spectacular and well-known atmospheric optical phenomenon is the rainbow. Rainbows are among the most remarkable effects in the atmosphere. Sunlight and water drops are necessary factors for the formation of the rainbow. Furthermore, the observer must be located between the Sun and rain. Since the colours of the spectrum are all bent at different angles, this refraction causes the colours to disperse or separate, as in a prism. The full spectrum of colours is then reflected off the back of the raindrop into the air. As the colours pass from the raindrop into the air, they are refracted for a second time. This second refraction causes the different bands of colours to become more distinct. Figure 21.7 shows a visual appearance of rainbow (left panel) and the schematic illustration how primary and secondary rainbows are formed due to light propagation in spherical droplets (right panel). The corresponding numbers denote spherical droplet (1), places where internal reflection of the light occurs (2), primary rainbow (3), places where refraction of the light occurs (4), secondary rainbow (5), incoming beams of white light (6), path of light contributing to primary rainbow (7), path of light contributing to secondary rainbow (8), observer (9), region forming the primary rainbow (10), region forming the secondary rainbow, (11) and zone in the atmosphere holding countless tiny spherical droplets (12). A rainbow is produced as the net result of this sequence of events repeated over and over when the atmosphere is filled with billions of tiny raindrops (such as after a storm). The curved form of a rainbow occurs because its rays always travel to the observer at an angle between 40° and 42° from the path of sunlight.

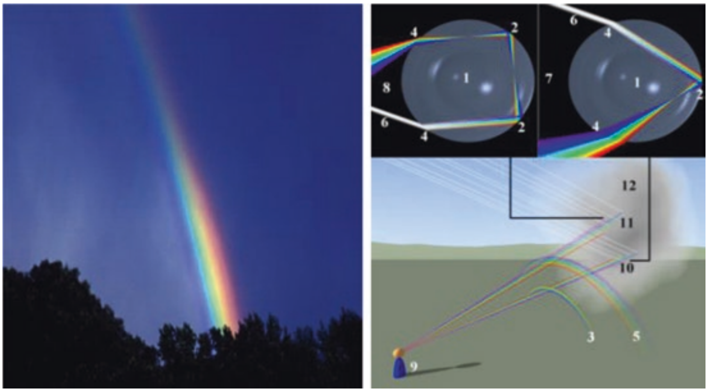


Fig. 21.7 Rainbow formation (Credit: Peo [CC BY-SA 3.0 (http://creativecommons.org/licenses/by-sa/3.0/)])

Fig. 21.8 Sun halo (Credit: Gabor Szilasi [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

21.2.3 Halo

Sunny halo refers on a narrow ring with a diameter of the sunset (Fig. 21.8). That usually happens when the sky is covered with a thin layer of cirrus clouds. The most common halo is the 22° halo, so named because its radius subtends an angle of 22° from the observer. Less observed is the growing 46° halo. 22° halo around the Sun and the Moon occurs because of refraction of light in thin hexagonal ice crystals in the air. The main difference between 22° and 46° halo is the path that takes the light through ice crystals. One of the most spectacular effects associated with Halo is *"Sun dogs"* or perihelia. These two clear regions, or "false-false Suns" as often called, can be seen near 22° Halo. Sun pillar often present near the sunrise or the sunset of the Sun, a vertical ray of light that appears to some range up from the sun. Set of optical phenomena in the form of a ring, which occurs with the reflection of sunlight or moonlight, the ice crystals that float in the atmosphere. Halo with all the accompanying phenomena is often seen at polar ends.

Observations showed that Halo is often seen in front of the cyclone and can serve as a sign of the emergence of the storm.

21.2.4 Glory

Glory is atmospheric optical phenomenon, noted by the pilots. This photometer consists of one or more colour rings surrounding the shadow of the observer (aircraft) projected on the clouds (Fig. 21.10). It is formed in a manner different from that of the rainbow.

Gloria can also be seen by observers in terms of fog on the Earth's surface when the sun is at its rear side.

21.2.5 Corona

Corona represents a series of colour rings with relatively small radiuses that are available through visible clouds around the Sun and the Moon (Fig. 21.11).

Solar corona is a unique optical phenomenon, often seen accompanied by the Moon and the Sun. This is clear photometeor whitening disk placed on the Moon or the Sun. The corona is created when water droplets in the thin layer of water clouds, usually altostratus have scattering light from the light body. Colours of the crown are the result of a process called diffraction, slight violation of the light when it passes through the cloudy droplets. Refracted light from all sides of cloudy drops will go into the shadow of the drops. Here light rays meet and interferes with the other rays that produce various components of white light, which generates colours that make a crown.

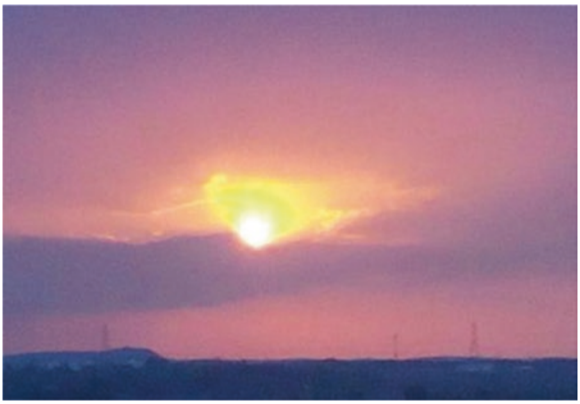


Fig. 21.10 The glory of the sunset (Credit: Shuvam swapnil [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

Fig. 21.11 Sun corona (Credit: Christos Doudoulakis [CC BY-SA 4.0 (https://creativecommons.org/licenses/by-sa/4.0)])

Fig. 21.12 Aurora borealis (Credit: Rafal Konieczny [CC BY-SA 3.0 (http://creativecommons.org/licenses/by-sa/3.0/)])

21.2.6 Aurora

Aurora is a natural light display in the Earth's sky mainly seen in the high-latitude regions (the Arctic and Antarctic). These beautiful optical effects are produced when the magnetosphere is sufficiently disturbed by the solar wind that the trajectories of charged particles in both solar winds, mainly in the form of electrons and protons, precipitate them into the upper atmosphere.

They are one of nature's most stunning displays, appearing as luminous streamers, arcs, curtains, or shells in the night sky (Fig. 21.12). Those that occur in the Northern Hemisphere are known as aurora borealis or northern lights; those in the Southern Hemisphere are known as aurora australis or southern lights. The energy particles at the Earth that form aurora come from aerospace middle-magnetosphere. These energetic particles are mostly electrons, but protons also produce aurora. Electrons travel along the lines of magnetic fields. Earth's magnetic field looks like a field of dipole magnet where the field lines appear and go into the Earth near the poles. Aurora's electrons in this way are referred to the high latitudes of the atmosphere. As they penetrate the upper atmosphere, the chance of collision with the atom or molecule increases as they go deeper. At the moment when the collision occurred, the atom or molecule takes part of the energy of the particle energy and keeps it as internal energy while the electron goes to a reduced speed. The process of preserving energy molecule or atom is called the "excitement" of the atom. An excited atom or molecule can return to non-exited state (initial state) by sending the photon out of each other, i.e. by creating light.

21.2.7 Colour of the Sky

Most of the molecules of the gas atmosphere contain oxygen and nitrogen. When sunlight strikes these molecules, the larger wavelengths of the light pass right through them, but the smaller wavelengths (violet, indigo, blue, and green) reflect (Fig. 21.13).When we look at the sky, only to see these scattered tapes, which appear as various shades of blue.

21.2.8 Cloud Iridescence

It is a colourful optical phenomenon that occurs in a cloud and appears in the general proximity of the Sun or Moon. Cloud iridescence represents beautiful optical phenomena (see Fig. 21.14). It is a type of photometeor, as a common optical effect that is most often observed in altocumulus, cirrocumulus, lenticular, and ice-crystal cirrus clouds that have brilliant spots of colours, usually red and green, observed up to about 30° from the Sun. Cloud iridescence is caused by water droplets diffracting light (within 10 degrees from the Sun and by interference effects. It can extend up to 40 degrees from the Sun. The cloud droplets are very thin (a few micrometres)

Fig. 21.13 The blue sky

Fig. 21.14 Cloud iridescence

and almost the similar size. They result from local adiabatic lifting and condensation process in moist air often in a lenticular (wave) cloud, above a developing cumulus, or occasionally in irregular spots of uniform colour in a region of shallow convection.

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Chapter 22
Atmospheric Chemistry

22.1 Units of Measurements

Atmospheric chemistry is an important scientific branch that studies the basic elements of the Earth's atmosphere, particularly the air pollution and its impacts (e.g. Warneck 2000; Finlayson-Pitts and Pitts 2000; Visconti 2016; Lamb and Verlinde 2011; Hobbs 2000; Bonan 2015). It specifically looks at the atmospheric chemical composition, the reactions, and interactions as driven mechanisms of the Earth-climate system. Atmospheric chemistry incorporates the significant topics as laboratory experiments, field measurements, and modelling. The main intention of this chapter is to introduce readers in basic elements of atmospheric chemistry, encompassing the chemical principles used to describe a complex system such as the Earth's atmosphere. According to Jacob (1999), one of the main objectives of atmospheric chemistry is to understand in general the chemical concept, measures, and factors that control the concentrations of chemical species in the atmosphere. Here we start with a brief description of basic units of measurements: mixing ratio, number density, and partial pressure.

22.1.1 Mixing Ratio

It is usually denoted as r_i of species (i) is defined as the ratio of the amount of the chemical specie in each volume to the total amount of all constituents in that volume or number of moles of r per mole of air:

$$r_i = \frac{\text{number moles } (i)}{\text{mole of air}} \tag{22.1}$$

V. Spiridonov, M. Ćurić, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_22

The units are volume of gas per volume of air or mol/mol. The mixing ratios of the permanent gases are as follows:

(N_2) with C_{N2} = 0.78 mole/mole; or N_2 or 78% of all molecules in the atmosphere
(O_2) with C_{O2} = 0.21 mole/mole or 21% of all molecules in the atmosphere
(Ar) with C_{Ar} = 0.0093 mole/mol

Mixing ratios of trace gases are commonly given in units of:

parts per million volume (ppm), where 1 ppmv = $1\text{x}10^{-6}$ mole/mole
parts per billion volume (ppb), where 1 ppbv = $1\text{x}10^{-9}$ mole/mole
parts per trillion volume (ppt), where 1 pptv = $1\text{x}10^{-12}$ mole/mol

22.1.2 Number Density

Since air is a mixture of gases, the volume mixing ratio of gas element (*i*) in the air r_{Vi} is given as the ratio between the number of molecules of gas (*i*), given as N_i divided by number of air molecules or unit volume of the atmosphere N_a, or usually given as a number of molecules of (*i*), per cm^3 of air:

$$r_{Vi} = \frac{N_i}{N_a} \tag{22.2}$$

This quantity is referred to us as a number density. It is an important measure for calculating gas-phase reaction rates or *column density* for measuring the absorption or light scattering by gases, given as the integral over the depth of the atmosphere:

$$r_{\text{col}\_i} = \int_{z}^{0} n_i \text{dz} \tag{22.3}$$

The number density and the mixing ratio of a given gas component are related by the number density of air n_a (molecules of air per cm^3 of air):

$$n_X = r_i n_a \tag{22.4}$$

Utilizing ideal gas law $pV = NRT$, with some replacements for the number density of air we get $n_a = \frac{A_V p}{RT}$ where A_v = 6.023x1023 molecules mol^{-1} is Avogadro's number. After substitution in (Eq. 22.4), we obtain:

$$r_i = \frac{A_V p}{RT} R_i \tag{22.5}$$

22.1.3 Partial Pressure

The partial pressure $\boldsymbol{p_x}$ of a gas i in a mixture of gases of total pressure p is defined as the pressure that would be exerted by the molecules of i if all the other gases were removed from the mixture. Dalton's law states that p_i is related to p by the mixing ratio R_i:

$$p_i = r_i \mathrm{p} \tag{22.6}$$

For our applications, p is the total atmospheric pressure. By making use the ideal gas law to relate p_i to n_i:

$$p_i = \frac{n_i}{A_V} \mathrm{RT} \tag{22.7}$$

The partial pressure of a gas determines the exchange rate of molecules between the gas phase and a coexistent condensed phase or aqueous-phase atmospheric chemistry. Concentrations of water vapour and other gases that are of most interest because of their phase changes are often given as partial pressures.

The absorption of a gas phase chemical species into the cloud water and rainwater is determined either by the equilibrium according to Henry's law or by mass transfer limitation calculations in order to include the possible non-equilibrium states.

Gases (with an effective Henrys law constant $\kappa_H^* < 10^3 moldm^{-3} atm^{-1}$) in cloud water and rain are assumed to be in equilibrium with the local gas-phase concentrations. These liquid-phase concentrations of each chemical component (i) are calculated according to Henry's law, i.e.:

$$[i] = K_H p_i \tag{22.8}$$

where [i] is given in units of $mol\frac{(i)}{L}H_2O$ or M, K_H is Henry's law coefficient ($Matm^{-1}$) and p_i is the partial pressure of the species [i] given in units atm. All equilibrium constants and oxidation reactions are temperature dependent according to van't-Hoff's relation:

$$K_T = K_{T0} \exp[-\frac{\Delta H}{R}\left(\frac{1}{T} - \frac{1}{T_0}\right) \tag{22.9}$$

where ΔH is the increase of enthalpy induced by chemical reactions, is the equilibrium constant at a standard temperature T_0 = 298 K, and R is the universal gas constant.

22.1.4 Dry and Wet Deposition

For many gases and for particles as well, deposition to surfaces competes with chemical reaction for the depletion of those species. We will consider dry and wet deposition. Dry deposition refers to the flux into surfaces in the absence of precipitation. Wet deposition of chemical species is associated with atmospheric moisture processes and applies to deposition into weather phenomena such as fogs, clouds, rain, or snow where the gas is incorporated in the bulk solution. In both cases, the relationship between the concentration of species (i) and the vertical flux, ϕ_i to a surface is given by:

$$\phi_i = V_d \left[C_i \right] \tag{22.10}$$

where V_d is deposition velocity of species (i) with units of, most commonly, cm/s. The deposition velocity is difficult to define by first principles, depending on landscape, the type of ground surface, vegetation, the atmospheric conditions (temperature, pressure, wind), and the reference height at which the concentration of species (i) is measured.

22.1.5 Gas-Particle Partitioning

Both, gas and particles can be a function of the solubility of the gas. Hence, an important parameter that determines the deposition velocity is the *Henry's law constant*, K_H, for the gas. Henry's law is the expression that defines the partitioning of a species between gas and solution phase:

$$\left[C_{sol} \right] = K_H \left[C_{gas} \right] \tag{22.11}$$

22.2 Chemical Composition of the Atmosphere

Atmosphere is an envelope of air surrounding the Earth and bound to the Earth by the Earth's gravitational attraction. It extends from the surface into space with decreasing density with height. For about 3/4 of the atmosphere by mass is in the lowest 10 km. About 99% of the mass of the atmosphere is located below 50 km, in the stratosphere and the troposphere. Gases other than N_2, O_2, A_r, and H_2O are present in the atmosphere at extremely low concentrations and are called *trace gases*. Despite their low concentrations, these trace gases can be of critical importance for the greenhouse effect, the ozone layer, smog, and other environmental issues. The bulk composition of the air (99.997% by volume) consists of mainly N_2, O_2, A_r, and

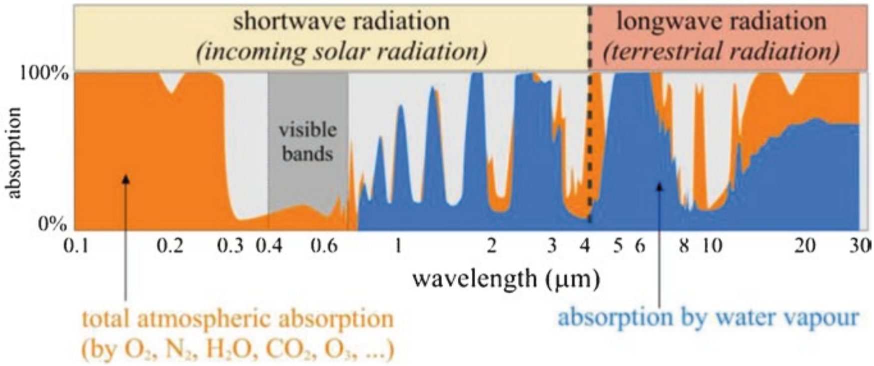

Fig. 22.1 Absorption rate in (%) of all atmospheric species in short and longwave radiation

CO_2. These are stable species with the little or no interesting chemistry. The remaining 1% of the atmospheric gases is known as *trace gases* because they are present in such small concentrations. CO_2 is a greenhouse gas (very effective absorber and emitter in the infrared). For example, O_3 is a good in stratosphere (protects from UV), but bad near the surface (corrosive to lungs and more).

SO_2 is emitted by volcanoes and participate in production of sulphate aerosols and acid rain, while NO_2 is produced by combustion and forms acid rain and smog. Figure 22.1 shows absorption of solar and terrestrial radiation by all atmospheric components and by water vapour.

22.2.1 Water Vapour as a Greenhouse Gas

Atmospheric water vapour has several significant direct and indirect effects on both weather and climate. It plays important roles in the radiation and the energy budgets of the atmosphere and in the formation of clouds and precipitations. About 70% of total absorption of the incoming shortwave solar radiation, particularly in the infrared region, and about 60% of total absorption of longwave radiation by the Earth are realized by water vapour; thereby it is the most significant greenhouse gas. Water vapour also influences heat energy transfer on the surface-atmosphere system through the latent heat flux. It plays an important role in the heat transfer from Earth's surface into the atmosphere. Relative concentration expressed through the mixing ratio as mole fraction C_X [mole mol^{-1}] remains constant when air density changes. Consequently, it represents a powerful measure of atmospheric composition. Air also contains variable H_2O vapour (10^{-6}–10^{-2} mole mol^{-1}) and aerosol particles.

$$r_i = \frac{number\ moles\ of\ i}{mole\ of\ air} \tag{22.12}$$

Table 22.1 lists the main atmospheric species and their mixing ratios. Trace gas (e.g. CO_2, N_e, O_3, H_e, CH_4, and K_r) concentrations are given in the following units:

$$1\,\text{ppm} = 1\,\mu\text{mole}\,\text{mol}^{-1} = 1\text{x}10^{-6}\,\text{mole}\,\text{mol}^{-1}\;1\,\text{ppb} = 1\,\text{nmol}\,\text{mol}^{-1} = 1\text{x}10^{-9}\,\text{mole}\,\text{mol}^{-1}$$

$$1\,\text{ppt} = 1\,\text{pmol}\,\text{mol}^{-1} = 1\,\text{x}\,10^{-12}\,\text{mole}\,\text{mol}^{-1}$$

The amount of water vapour in the atmosphere is (~ 0 to 4%). It is critically important variable gas for clouds and precipitation formation and important way to move energy around. Water vapour and methane (CH_4) are major greenhouse gases. There are also non-gaseous constituents in the atmosphere:

- Hydrometeors-rain clouds, hail.
- Particulates and aerosols.
- Aerosol is a liquid or solid dispersed in a gas, usually air.
- Particulates which can be inorganic (soil, smoke, dirt, sea salt, volcanic dust, surface acid aerosol) or organic (seeds, spores, pollen, bacteria).

The particles in the air are important as they act as condensation and freezing nuclei. Water likes to condense on/or freeze on to particles. They can absorb or scatter radiation, reduce visibility, and scatter solar radiation to space causing cooling of our Earth's planet. The particles can also impact human health through irritation of lungs and initiate asthma and heart disease.

22.3 Ozone Photochemistry

The layer above the troposphere which lies between 10 and 50 km above the Earth surface is called stratosphere (ozonosphere). As the result of absorption of solar radiation by the stratospheric ozone layer, air temperature is constant or higher with

Table 22.1 Atmospheric species

Species	Mixing ratio (dry air) [mol mol^{-1}]
Nitrogen (N_2)	0.78
Oxygen (O_2)	0.21
Argon (Ar)	0.0093
Carbon dioxide (CO_2)	400×10^{-6}
Neon (Ne)	18×10^{-6}
Ozone (O_3)	$(0.01–10) \times 10^{-6}$
Helium (He)	5.2×10^{-6}
Methane (CH_4)	1.8×10^{-6}
Krypton (Kr)	1.1×10^{-6}

height. It is mainly stable (not a lot of vertical mixing) and dry layer. Only occasionally get overshooting tops from convection pushing into this layer. Maximum ozone (O_3) is found in the layers between 20 and 30 km. Ozonosphere protects the surface from harmful ultraviolet (UV) radiation. Ozone is produced photochemical. A photochemical reaction is a chemical reaction in which radiation affects the chemical composition of a substance. The first slowly moving process is photolysis (separation of molecules by the action of light) under influence of the Sun, when oxygen molecules are *photolysis*, yielding one oxygen atoms. The process follows by mutual conversion of the ozone and oxygen atoms to solar UV breaks ozone. During this fast process, the oxygen atom reacts with another oxygen molecule. This interconversion processes transforms UV radiation into thermal energy, heating the stratosphere. The ozone layer is monitored both by satellites and ground-based instruments that are dedicated to observing the destruction of stratospheric ozone. The TOMS satellite measures the ozone levels using the back-scattered sunlight method in the ultraviolet (UV) range (Fig. 22.2). Thickness of ozone layer is measured as a column concentration expressed in "Dobson Unit (DU)" = 2.69×10^{16} molecules cm^{-2}.

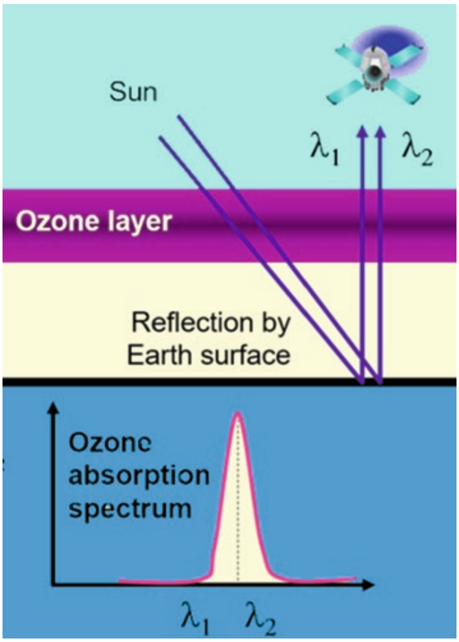

Fig. 22.2 The back-scattered sunlight method for ozone monitoring

22.4 Atmospheric Pollution

Air pollution is a condition when pollutants under certain atmospheric conditions are present in concentrations high enough to affect adversely on human health and cause environmental damage. Air pollution causes acid rain, ozone depletion, photochemical smog, and other similar phenomena. The substances include natural or anthropogenic chemicals that may exist in the atmosphere in gas, liquid, or solid state. Atmospheric conditions affect the transport, dispersion, and redistribution of deposition of pollutants. On the other hand, air pollution affects weather and climate. The air is never perfectly clean.

22.4.1 The Mechanisms of Air Pollution

Atmospheric pollutants are gaseous or solid particles present in air, which have a detrimental effect on human health and endanger the natural environment. Pollutants which have significant harmful effects on humans and the environment are:

- Presence of inert dust
- Local pollutants in the air
- Toxic substances in air, radiation
- Photochemical smog in urban areas
- Acid rain
- Reducing visibility
- Warming caused by greenhouse effect
- Distortion of the ozone layer
- Climatic forcing due to anthropogenic factors
- Atmospheric aerosols

Air pollution consists of three main components: emission sources, medium (atmosphere), and receptors. Basic sources of emissions are transportation, industrial, and domestic fuel combustion and industrial processes. Key receptors are humans, animals, plants, and materials. The atmosphere in this system participates as a medium for transport and dispersion of the physical and chemical transformation of pollutants.

22.4.2 Classification of Pollutants

According to chemical composition, air pollution can occur as a result of the presence and activity of the following chemical species:

- Sulphur compounds
- Nitrogen compounds

- Carbon compounds
- Halogen compounds
- Toxic substances
- Radioactive compounds

In terms of physical condition, atmospheric pollutants are divided into gas, liquid, and solid pollutants. According to the way pollutants are present in the atmosphere, they can be grouped into two categories:

1. Primary pollutants, which are emitted directly from existing sources
2. Secondary pollutants that create the chemical reactions between primary pollutants and normal atmospheric conditions

Primary pollutants include particulate matter (PM), sulphur dioxide, nitrogen oxides, volatile organic components (VOCs), carbon monoxide, and lead. Secondary pollutants are atmospheric sulphuric acid air pollution in urban and industrial areas known as smog, photochemical smog which is a harmful mixture of gases and particles. It occurs under the influence of strong sunlight, by some gas activated molecules or atoms of gas, and they look like photochemical reactions. The main component of smog is ozone. According to the spatial scale, atmospheric pollutants are divided into local, regional, or global. The Environmental Protection Agency (EPA) defined six (6) main pollutants which set ambient air standards in order to protect the health of people and material goods. Measurable pollutants (defined by EPA) are:

1. Ozone ($O_{3)}$
2. Carbon monoxide (CO)
3. Sulphur dioxide (CO_2)
4. Nitrogen oxides (NO_x)
5. Lead (P_b)
6. Particulate matters (PM_{10} and $PM_{2.5}$)

22.4.3 Primary Atmospheric Pollutants

Ozone (O_3) is a reactive gas with no colour and no smell and is composed of three oxygen atoms. It is naturally found in the Earth's stratosphere, where it absorbs ultraviolet radiation, which can be harmful to life on Earth. Ground ozone is "harmful", and it is a major constituent of photochemical smog. Basic pollutants involved in these reactions are nitrogen oxides (NO_x) and volatile organic components (VOCs). Carbon monoxide (CO) also participates in the reactions and supports the process of formation of ozone. Warm and sunny conditions during the calm and stagnant wind are suitable for forming ozone.

Sulphur (S) is one of the most reactive chemical elements. Pure sulphur is a nonmetallic element with pale yellow, weak conductor of electricity and is insoluble in water. It reacts with all metals except gold and platinum, forming sulphides. It also forms compounds with several non-metals. Each year millions of tons of sulphur are produced, mostly in the manufacture of sulphuric acid, which is widely used in industry. Natural sources of sulphur are marine plankton, sea water, bacteria, plants, and volcanic eruptions. Sulphur dioxide ($SO_{2)}$ is a colourless gas with a sharp smell, a basic pollutant that has anthropogenic and natural sources. Anthropogenic sources are industries that burn fossil fuels containing sulphur, lignite mines, power plants, and oil refineries. At relatively high concentrations (SO_2), they cause dangerous respiratory problems. Sulphur dioxide is a precursor of sulphuric acid which is a source of acid rain that is created when (SO_2) combines with water droplets. Sulphur dioxide is a precursor of sulphate particles (sulphates) that affect radiation balance in the atmosphere and could cause global cooling.

Nitrogen (N_2) is the dominant gas in the atmosphere, which takes about (78%) by volume. Nitrogen oxide (NO_X) is found in the non-sustainable (non-determined) mixture of nitric oxide (NO) and nitrogen dioxide ($NO_{2).}$ Nitrogen oxides (NO_X) are mainly composed of (N_2 and O_2) at high temperatures, as a result of combustion of fuel in vehicles. Natural sources are a certain kind of bacteria, electrical discharges, and combustion of biomass. Nitrogen oxides cause reddish-brown haze in urban air, which contributes to the increased occurrence of heart and respiratory diseases. Nitrogen oxide ($NO_{2)}$ is the precursor of acid as a source of acid rain, which is created when nitrogen oxides combines with water to create nitric acid. Nitrogen oxides are precursors of nitrogen suspended particles (nitrates) that affect the energy balance of the atmosphere. Nitrogen oxides (NO_X) separately (N_2O) destroy the ozone layer. This layer absorbs ultraviolet light, which is potentially dangerous to life. Nitrogen oxides are major factors for the formation of ground floor "bad" ozone.

Carbon (C) is a naturally rich non-metallic element found in many different compounds: food, clothing, cosmetics, fuels, and others. Carbon is the sixth most important element in the universe and plays a crucial role in the chemistry of life. Carbon monoxide (CO) is colourless, odourless, flammable gas, and a major pollutant in urban air, which creates the incomplete combustion. Anthropogenic sources are motor vehicles with gasoline engines, smoke cigarettes, burning of biomass, while natural sources of carbon monoxide arise from combustion of biomass. Carbon monoxide (CO) is also produced by atmospheric oxidation of methane gas and other hydrocarbons. Carbon monoxide is highly poisonous to humans and animals. When inhaled (CO) reduces the ability of haemoglobin binds to oxygen in the blood. Carbon dioxide is the complete oxidation product of burning fuel. Also, in the atmosphere, (CO) is present in the oxidized (CO_2). Carbon dioxide (CO_2) is a key greenhouse gas. Primary sources include combustion of fossil fuels, cutting forests, and cement production. U.S. are the largest single emitters of (CO_2), which contribute to about (16%) of the total global amount.

Volatile hydrocarbons (VOCs) are those gases that contain hydrogen and carbon but may also contain other atoms. Hydrocarbons (HCs) are organic gases containing only carbon and hydrogen. Volatile organic components (VOCs) are hydrocarbons that contain methane and oxidized hydrocarbons (hydrocarbons with oxidation groups). Methane (CH_4) is the most present hydrocarbon in the atmosphere, which is in the exhaust gases from cars, the combustion of biomass in agricultural activities (e.g. rise fields). Anthropogenic sources are combustion of fossil fuel and evaporation of gasoline (e.g. oil refineries during the spending on fuel vehicles). Natural resources (HCs) are the dissolution of organic matter from certain types of pine forest, shrubs, and the like. Some hydrocarbons (formaldehyde) pollutants are only local, but others contribute to smog containing ozone.

Chlorofluorocarbons (CFCs) are artificial gases, which are used as coolers in refrigerators and cooling systems (air-conditions). They are neither toxic nor flammable. The most present chlorofluorocarbons are ($CFCl_3$) and (CF_2Cl_2). CFCs are artificial halocarbons; hence they are not biologically degraded. Also, chlorofluorocarbons are not soluble in water, so they are scavenged from the atmosphere by rainfall. In the stratosphere, UV destroys (CFCs), degrading them to several chemicals (including chlorine atoms and bromine atoms, which effectively destroy ozone). CFCs are the main greenhouse components and lead to reduction of stratospheric ozone.

Metals such as mercury, cadmium, chromium, and nickel lead represent one type of contaminants found in impurities and fuels. Anthropogenic sources of metals are mining emissions of metals from industrial facilities, motor vehicles, and others. For example, lead is a very useful metal, which has been mined for several thousand years, but this and other metals are highly toxic.

Particulate Matters (PM_{10} and $PM_{2.5}$) are solid or liquid particles, composed of one or more chemicals, which are small enough to remain suspended in the air. Examples of particulate particles are inert dust, smoke, smoke, sulphates, nitrates, asbestos, pesticides, bio-aerosols (e.g. pollen, spores, bacteria, fragments of insects, etc.). Particulate matters ($PM_{10)}$ are particles with a diameter (<10 μm), while $PM_{2.5}$ is particles with a diameter < 2.5 μm. Anthropogenic sources of particulate matters are burning of biomass, conversion of the particulate emissions, industrial processes, and agricultural activities. Natural sources include sea salt, sandstorms, burning of biomass, conversion of gases to particles, and volcanic wastes. Particulate matters have negative impacts on health and cause much damage to the human respiratory system, also contributing to the emergence of haze in urban areas, causing a reduction in visibility.

22.4.4 Factors Affecting Air Pollution

Most obvious factor affecting the air pollution is the quantity of contaminants emitted into the atmosphere. However, certain episodes of pollution in the air do not occur because of the large emission of pollutants in the atmosphere, but due to changing atmospheric conditions that affect the transport, dispersion, and the redistribution of pollutants. Work parameters that define the atmospheric conditions are the stability of air disturbances, mixing of air, humidity, wind speed, wind shearing and veering, etc.

Air Mixing Direct effect of the wind speed is in the influence of the concentration of pollutants. Atmospheric stability determines the extent to which vertical movements will be mixed with clean air pollution, which lies above the surface layers. Vertical distance between the Earth's surface and the height to which it is manifested is stretching through convective movements called thickness (layer) of the interference. Generally, the greater the mixing layer is, the better the air quality.

Thermal Inversion The temperature inversion is a condition in which the temperature gradient in a layer of the atmosphere is negative. It means that if the conditions are standard, the temperature decreases with height, and inversely, temperature increases with height. In terms of inversion, the atmosphere is stable, and an energy barrier at the ground layer of the atmosphere. In areas where there are sources of air, pollution is stagnant with no wind, and we should expect existence of high concentrations. Upper level temperature inversions are associated with lowering the air, which is characteristic of the centres of high pressure (anticyclones).

22.4.5 Acid Rain

Acid rain in the wider sense indicates a mixture of wet and dry deposition from the atmosphere, which contains larger quantities than normal amounts of nitric and sulphuric acid. Predecessors in creating acid rain originate from natural sources such as volcanoes and the dissolution of vegetation and from anthropogenic sources, primarily emissions of sulphur dioxide (CO_2) and nitrogen oxides (NO_X) resulting from the combustion of fossil fuels. In the United States, nearly 2/3 of the total (CO_2) and 1/4 in (NO_X) comes in electricity production, which relies on burning fossil fuels such as coal. Acid rain is occurring when these gases in the atmosphere react with water, oxygen, and other chemicals to form various acidic compounds. The result is a mild solution of sulphuric acid and nitric acid. When sulphur dioxide and nitrogen oxides are released from power stations and other sources, prevailing winds wash over large distances. The value of the pH factor of precipitation in industrial areas and in urban areas where there are activities by people is much larger than when it was registered in non-urban and rural areas. These acidic rain or snow is formed when sulphur or nitrogen oxides produced as ancillary products of combustion and industrial activity are converted to acid during complex atmospheric reactions. By reducing pH value of precipitation, the harmful effects on human health and natural environment are increased. Acid rain is toxic to fish; they contribute to change in complex ecosystems, with many interactions of many levels in the organization.

Wet deposition refers to acidic rain, fog, and snow. If the acid chemicals in the air lie in areas where the weather is wet, acids can fall to the Earth's surface in the form of rain, snow, or fog and adversely effect on plants and wildlife.

Dry deposition. In areas where the weather is dry, some acidic chemicals can bat of dust or smoke and fall to Earth via dry deposition. Dry deposited gases and particles can be scavenged from these surfaces by storms, which lead to increased surface runoff.

Source emissions. Primary sources of emissions can be:

- Biological
- Ocean
- Solid ground
- Industrial
- Photochemical

Anthropogenic sources of emissions are fossil fuel and combustion of biomass. Biological sources of emissions are photosynthesis, biological sources of methane, nitrogen oxides (nitrification and denitrification bacteria in the land), and plants (isoprene-reactive hydrocarbons, etc.). In biological activities in the oceans different chemicals are released: (DMS, H_2S, CH_3Cl, CH_2S) including hydrocarbons. The following chemical compounds are released from the rigid ground of the volcanoes: SO_2, H_2S, H_2O, CO_2, HCl, and other mixtures. Other sources of emissions are the emissions of fossil fuels, caused by anthropogenic influences in the Northern Hemisphere, and emissions of combustion of biomass in tropics.

Transport of gases. The transport of gases on the planet Earth has different temporal and spatial scales. Typical time scale of meridional transport distinguishes up to 2 weeks, while the zonal transport takes (1–2) months to 1 year. Ground transport of gases from the surface of the Earth to the top of the planetary boundary layer (PBL), which is the amount of (1–3) km, is (1–2) days, while the higher layers of the atmosphere to the ground floor layers of the atmosphere in the troposphere is 1 week, and the beginning of tropopause time scale is 1 month.

22.5 Atmospheric Aerosols

Aerosols are airborne particles and/or liquid droplets and gases together suspended in the air. Under mean atmospheric aerosols are dispersed condensed particles in the air, usually spherical in shape, whose dimensions are in the interval of 0.001–100 μm (small drops). Fine airborne particles have a typical diameter in the range of 0.01 μm −100 μm. Marine and continental aerosols belong in the range 0.1–1.0 μm. Basically,

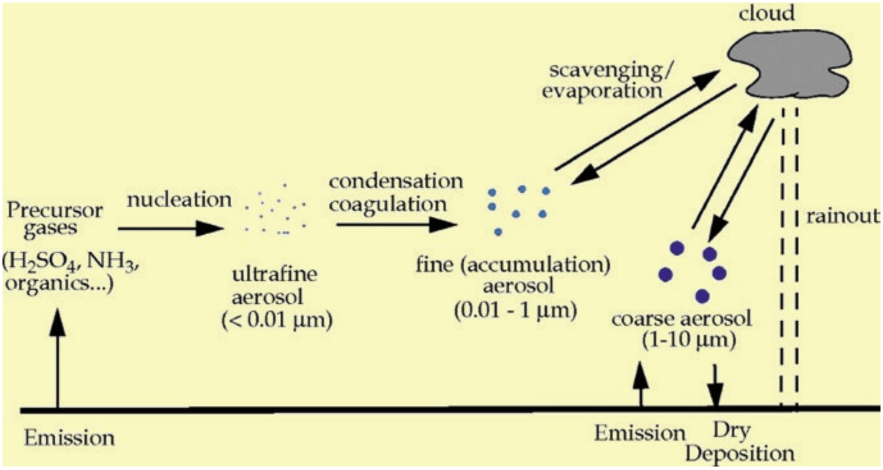

Fig. 22.3 Schematic illustration of the source of atmospheric aerosols

aerosol particles originate from the fires in forest and homes, industrial activity, biomass burning, sea spray, including volcanic eruptions, dust storms, living vegetation, and sea waves as natural sources. Human activities like burning of fossil fuels and the alteration of natural surface cover also generate aerosols. Average world aerosols generated by human activities currently contribute to about 10% of the total amount of aerosols in the atmosphere. Most of these 10 percent are concentrated in the northern hemisphere, particularly in the direction of the wind in industrial facilities and agricultural areas where they cut and burn forests and grass hair. Schematic representation of spring and the formation of atmospheric aerosols are given in Fig. 22.3. Scientists have much to study about the way aerosols affect regional and global climate (Seinfeld and Pandis 2016). Necessary additional knowledge to accurately determine the relative impacts of climate change by natural aerosols and those of human origin. Furthermore, insufficient information on which regions of the planet, the number of aerosols in the atmosphere increases, which decreases, and where the reserves are approximately constant. Aerosols tend to cause cooling of the Earth's surface. Because most aerosols reflect sunlight back into space, they have a direct effect of cooling by reducing the amount of solar radiation coming to Earth's surface.

The size of this cooling effect depends on the size and composition of aerosol particles and the basic reflective properties of the surface. It is assumed that the cooling effect of aerosols may partly be compensated by the expected global warming, which is associated with increasing the amount of carbon dioxide caused by human activity.

22.5.1 Effects of Atmospheric Aerosols

Atmospheric aerosol particles play important role in radiation budget of the Earth's as they scatter and absorb both shortwave solar radiation and longwave terrestrial radiation. They are also highly involved in the formation of clouds and precipitation since they operate as cloud condensation and ice nuclei (CCN and IN). Aerosol particles in the upper atmosphere, where the major part of atmospheric ozone forms, can modify the ozone removal. Additionally, particles are major elements of lower tropospheric air quality and can influence harmfully the environment and human health. Aerosols play important role in the balance of the Earth's climate. Due to the increasing anthropogenic emission of aerosols since the industrial revolution, they can also affect the global climate change. However, the effects of aerosols on climate are not one-way, moreover excessively uncertain. The climate forcing by aerosols can be realized through direct (scattering radiation) and indirect (cloud formation effects) radiative forcing.

22.5.2 Origin of Aerosols

There are three hypotheses about the origin of these particles: cosmic, continental, and oceanic. Regardless of the fact where it came from the atmosphere as has occurred (the natural or human activity), these atmospheric particles called aerosols. Aerosols have a different function in the atmosphere, and thus they have additional names. Cosmic dust comes directly from the cosmos, or the combustion occurs with some smaller celestial bodies (meteorites). These particles have a small percentage contribution to the atmosphere, compared with solid particles of different origin. The calculations show that the day of the cosmos in the atmosphere reach about 1000 tons of this dust. Continents represent a significant source of non-hygroscopic and hygroscopic particles that reach the atmosphere by the winds. It especially occurs in desert and dry areas. It is estimated that from North Africa to Europe, with currents are transported around 3.5 million tons of desert dust. A major hygroscopic source of particles is the process of combustion and rot of various organic substances, which are mainly in the form of sulphur and nitrogen compounds. In recent times, in urban and industrial areas large amounts of sulphur are released, for the combustion of coal, which contains from 0.3 to 3% sulphur. The released sulphur oxidation is passing in non-hygroscopic sulphur dioxide ($CO_{2)}$. Under the influence of solar radiation, oxidation continues and hygroscopic SO_3 is formed. With its merger with water vapour, sulphuric acid occurs. Volcanic eruptions appear as natural sources of hygroscopic substances on the mainland, which in average do not represent significant sources of aerosols, but mighty volcanic eruptions. Very important source of atmospheric drops is dispersion of small droplets from worldwide ocean areas and forming aerosols. According to the Aitkin's theory, the presence of sea salt in the air is explained by wave disturbances on the ocean surface under the influence of the winds. The waves are violated in the air and reach a huge number of tiny drops of seawater. During evaporation, small floating microscopic particles of salt remain of these drops. Turbulent movement of air and the wind lets them spread in the atmosphere and distribute large distances to several hundred kilometres. Measuring the size of salt particles in the atmosphere shows that the lower limit of its diameter is 0.2 μm. It is estimated that the mean concentration in cores of marine salt over the area of violations of the waves is (100/cm^3), with average production of sea salt formation (100/cm^2 s).

22.5.3 Aerosol Particle Concentration

One of the oldest and most suitable devices for measuring concentration atmospheric aerosols is an Aitken counter. This instrument expands the air with saturated water vapour, so that the water vapour becomes saturated by several hundred percent with respect to water. At such high saturation, water condenses on virtually all aerosol particles and a cloud of small droplets forms. The concentration of droplets in the cloud (close to the aerosol concentration) can be found in such a way that the droplets will precipitate on the substrate. They can then be viewed with a microscope or using another optical technique. The aerosol particle concentration measured by the Aitken counter primarily refers to the number of Aitken aerosol particles to be discussed later. The concentration of Aitken aerosols observed at ground level depends on the location. Likewise, changing the concentration of Aitken aerosols over time in one location can be up to more than one order of magnitude. In general, average aerosol particle concentrations are 109 m − 3 above the ocean, 1010 m^{-3} above rural areas, and up to 1011 m^{-3} above polluted urban areas. The measurements also show that the particle concentrations of Aitken aerosols decrease with height (Fig. 22.4). This is a result of the fact that soil is an important source of aerosols and especially human and industrial activity.

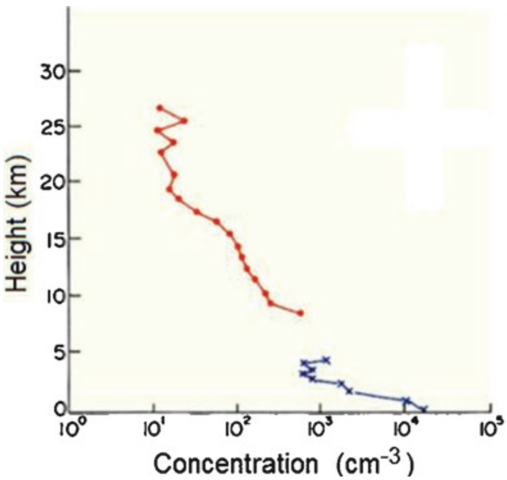

Fig. 22.4 A typical change in the concentration of Aitken aerosols with altitude. The upper curve obtained from the balloon and the lower curve from the plane measurements

22.5.4 Aerosol Particle Size Distribution

Aerosol particles are present at high concentrations in the atmosphere. Therefore, it is important to know the mean characteristics of the entire aerosol population, not each particle individually. In this connection, it is convenient to assume that the aerosol particles are spherical in shape. The magnitudes of atmospheric aerosol spherical particles range from 10^{-4} to tens of micrometres in diameter. Depending on the particle size as well as the measurement location, aerosol concentrations vary from 1013 to 1 m^{-3}. Because aerosol particle sizes are very different, different measurement techniques must be used to measure their characteristics.

So, aerosols with diameters between 0.003 and 1 μm can be measured by an electrical analyser. First, the particles are charged with a known charge and then collected in a certain way by the action of an electric field. The variation in magnitude of the collected charge is interpreted by the corresponding distributions of aerosol particles by the size of the diameter. Aerosol sizes of 0.3 to 30 μm can be determined by measuring the amount of light energy they dissipate. Aerosol it can also be collected on different surfaces and examined by optical or electron microscopes. Larger aerosols (> 1 μm) can be detected by various analytical techniques (X-ray energy dispersive analysis). In order to determine the concentration of aerosols collected, the efficiency of aerosol collection must be known. Larger aerosols are collected more efficiently than smaller ones. Therefore, only aerosol concentrations greater than 0.1 μm can be measured by direct collection. Average aerosol concentrations depending on its size for continental, marine, and urban polluted air are shown in Fig.22.5. The concentration of aerosol particles of diameter D on the ordinate is shown over the magnitude of dN / d (logD) and on the abscissa is logD. Size N is the concentration of aerosol particles larger than D. From Fig. 22.5, the following conclusions can be drawn:

- The concentration of aerosol particles decreases rapidly with the increase of their particles size. Therefore, aerosol particles with a diameter have the highest concentration less than 0.2 μm. These aerosols are called Aitken aerosols;
- Parts of the curves shown can be well described analytically as:

$$\frac{d\mathrm{N}}{d\left(\log\mathrm{D}\right)} = \mathrm{CD}^{-\beta},$$

  where C is a constant relating to aerosol concentrations and β is the slope of the curve between 2 and 4. A continental aerosol with diameters greater than 0.2 μm is well described by the preceding term when β = 3:

- Observed aerosol concentration distributions agree with measurements

  obtained by the Aitken counter, which indicate that the concentrations of aerosol particles are, on average, highest above the polluted urban areas and at least in sea environment:

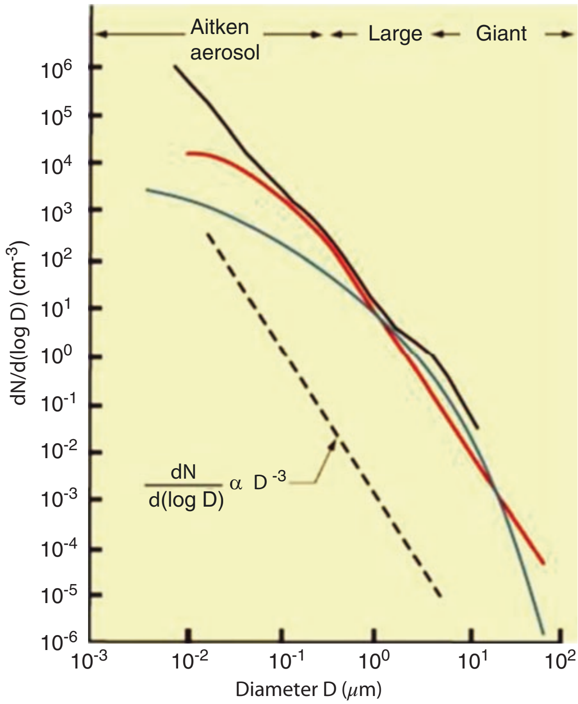

Fig. 22.5 Aerosol particle number distributions by size, determined by averaging datasets taken in marine air (blue line), continental air (red line), and urban polluted air (black line). A separate point-line-point line was obtained from the distribution of aerosol particles by β = 3

– Concentrations of aerosols with diameters greater than 2 μm (so-called giant aerosols) aerosols) do not differ so much for continental, marine or air above polluted urban areas.

The total concentrations of aerosols are on average largest in urban polluted areas and smallest in clean marine air. Typical values of the concentration of aerosols are (103/cm^3) over the oceans (104 /cm^3) than rural areas and land (105/cm^3) in polluted areas. The concentration of aerosols changes in space and time. Concentrations are usually highest near the Earth and near the sources. The concentration of aerosols decreases with increasing altitude. Small ions have almost no role in atmospheric condensation due to very high saturation necessary for condensation. The basic particles are only able to stay in the air for a limited time. Aerosols have a wide range of concentrations that vary from (10^7–10^{-6} cm^{-3}). Concentrations of aerosols decline very rapidly with the increase of their dimension. Atmospheric aerosols are divided into three main groups according to their dimension:

1. Aitken's particles = > 0.01 μm < D μm 0.1 μm
2. Large nucleus = > 0.2 μ m < D μm 2.5 μm
3. Giant nucleus = > D > 2.5 μm

22.5.5 Sources of Atmospheric Aerosols

Primary sources of atmospheric aerosols are:

- Condensation and sublimation of the vapour and the formation plumes at natural and anthropogenic combustion
- Reactions between tracers, i.e. gases in the form of traces (hydroxyl radical (OH), reactive nitrogen substances, hydrocarbons, ozone, sulphur components) into the atmosphere through the action of heat, radiation, or moisture
- Mechanical destruction and dispersion of matter on the Earth's surface, or sea sprayer over oceans, or as mineral dust over the continents
- Coagulation of the nuclei, which tend to produce large particles of mixed composition

Typical substances formed in large amounts of condensation Mon combustion, the ash, smoke from coal-tar products, oil, and sulphuric acid and sulphates. These particles are primarily within the Aitkin nuclei. Mechanical breakdown by wind and water on the rocks and land creates particles with diameters >0.2 μm. These primarily fall into the class of large nucleus (nuclei). Approximately 25% of the number of particles with a diameter greater than 0.2 μm is biological. Chemical reactions between nitrogen, oxygen, water vapour, and various tracers (e.g. sulphur dioxide, chlorine, ammonium, ozone, and oxides of nitrogen) primarily produce particles in Aitkin mode and greater. The examples are the formation of the ammonium chloride (NH_3 and HCl), oxidation of (CO_2 in $H_2SO_{4)}$ reaction of sulphur dioxide, ammonium, and water to form particles of ammonium sulphate (Spiridonov and Ćurić 2005) and the production of higher oxides of nitrogen by the action of heat, ozone, and ultraviolet radiation.

22.5.6 The Role of Clouds in the Atmosphere Pollution

The clouds have a major role in the process of transport and transformation of atmospheric pollutants. Transport and scavenging of chemical constituents, especially in convective clouds. Clouds with strong vertical development are important for understanding the composition of the troposphere and hence issues related to climate and chemistry of air quality.

Convective process is a very important mechanism for rapid transfer of chemical compounds and gases in very small concentrations (tracers), the border between layer and free troposphere, and is often an effective way of cleaning the atmosphere by wet deposition. Figure 22.6 shows a three-dimensional simulation of transport, redistribution and production of sulphates in convective clouds. The effect of convection on chemical substances in these clouds is critical to understanding the chemical-climate, air quality studies, and effects of acid rain on the Earth's surface.

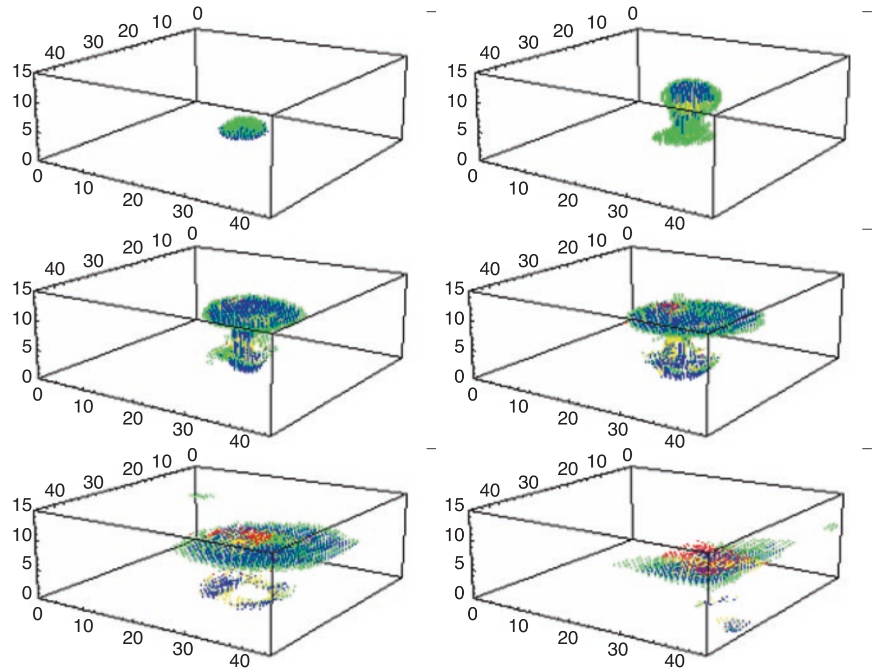

Fig. 22.6 Aqueous sulphate aerosol chemistry (three-dimensional simulation of transport, redistribution and production of sulphate aerosols in convective clouds. Blue, green, yellow, and red areas are concentrations of sulphate, hydrogen peroxide, sulphur dioxide and ozone, Spiridonov and Ćurić 2005))

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Chapter 23
Weather Forecast and NWP

23.1 Weather Forecast Overview

Weather forecast is a branch of the applied meteorology which studies the methods and techniques for prediction of the future state of the atmosphere at a given location or an assessment of the future behaviour of the atmosphere with respect to precipitation, clouds, winds, and temperature (Lazić 2010). Weather forecasting is one of the important science applications in our day-to-day planning activities. This is one prominent application that has played a significant role to humankind from long way back (Thaxter 1990). Wherever humans have settled around the world, weather forecasting has always been part of their life for man which has always been actively involved in their surroundings. Such assessments are usually made by government or private meteorologists, often using numerical simulations. Such simulations are the result of representing the atmosphere mathematically as a fluid in motion. This was done through collecting of quantitative data about the current state of the atmosphere through scientific understanding of atmospheric processes and numerical calculations of the equations which best describe the atmospheric processes. Due to the waves and the chaotic nature of the atmosphere, enormous computing power is required to solve the equations that describe the behaviour of the atmosphere.

23.2 Historical Background

Every night billion people worldwide monitor the weather forecast. What will the morning bring? After we listen to the forecast, we plan our daily activities. But how has the weather forecast been developed during the past years? Weather forecast in Biblical times was based on singular observations of the sky. Weather forecasting has become an important field of research and has always been done since the

V. Spiridonov, M. Ćurić, *Fundamentals of Meteorology*,
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ancient times, although the trend, methods, and techniques have changed with time. It began with the early civilizations of humankind using their personal experience to monitor seasonal changes and other reoccurring meteorological events in the atmosphere. Since ancient times, people have been trying to predict the future behaviour of the atmosphere. Here are thoughts from the Biblical book of Matthew, where Jesus says to the religious leaders of first century: "You're able to interpret the appearance of the sky, but anticipation of weather you cannot interpret". This method of simple observation dominated until 1643 when the Italian physicist Evangelista Torricelli discovered Barometer.

23.2.1 Early Research Period

As highlighted earlier, weather forecasting began with the early humankind using reoccurring astronomical and meteorological events to monitor seasonal changes in the atmosphere. For example, the Babylonians predicted short-term weather changes based on cloud appearance and optical phenomena; the Chinese divided the year into 24 festivals – each associated with a certain type of weather phenomena; the Greeks developed theories about rain, cloud, lightning, and other observations. A notable example is Aristotle who wrote in his Meteorologica (Fig. 23.1) a philosophical treatise including theories about the formation of rain, clouds, hail, wind, thunder, lightning, and hurricanes. He made some remarkable observations concerning weather. For almost two millennia, meteorological thought was based on

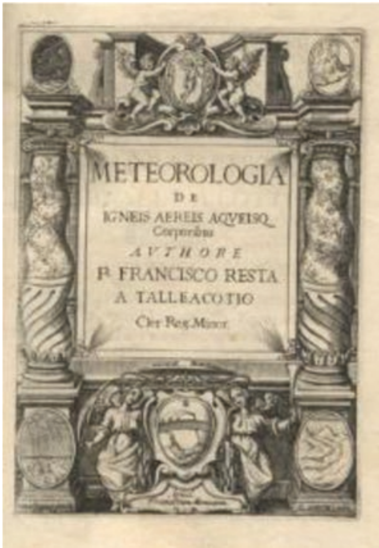

Fig. 23.1 The Aristotle's meteorologica

the ideas set out by Aristotle. Aristotle's ideas dominated intellectual thought for centuries, although it was a product of observation rather than experimentation. Weather forecasting was not considered as a science at the time; there was no experimentation or theoretical testing. It was more like an art than a science. Weather forecasting has always played a vital role in peoples' lives in their everyday activities. With the forecast information, people can know and plan better for what would come or to expect. For example, hunters, farmers, fishermen, warriors, shepherds, and sailors needed to know how the weather might be in the next day, because they worked in the open. Understanding weather patterns and atmospheric changes has many applications across various sectors and societal importance: warning for/about severe weather, agriculture, transport services, flood warning, commerce, industry, etc. For example, the growing vulnerability of densely populated areas and natural hazards makes increased demands for reliable forecasts of the consequences for the safety of life and property. Pilots need to plan for their flights; sailors need to plan their [marine] activities; farmers need to plan for watering. Important parameters to know include temperature, wind speed and direction, clouds, precipitation, visibility, humidity, and/or trends in all the above. The timing of significant changes and occurrences of extreme events is very important. Although attempts were made, safety against bad weather events, for example, was never guaranteed. An estimated 459 people drowned in the island near Wales in 1859, when the Royal Charter gold ship was wrecked in a large storm off Anglesey. This havoc influenced Robert Fitzroy (1805–1865: an English sailor, royal navy and scientist) to develop weather charts (Fig. 23.2) to allow forecasts to be made to "improve safety" at the sea. For example, he organized for ship captains to be provided with weather instruments in order to collect data (e.g. barometric temperature, air and sea temperature, humidity, wind, cloud, and soundings) and return it for computation.

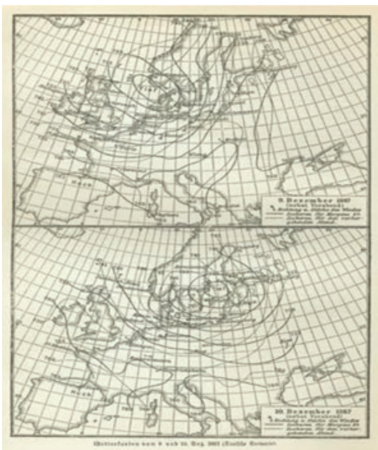

Fig. 23.2 The first weather charts

23.2.2 Weather Forecast Evolution

Weather forecasting was pioneered by Robert Fitzroy in the mid-1800. He developed fundamental techniques for weather forecasting and put a lot of effort in studying the weather, using his own nautical charts. He applied a qualitative knowledge of the atmosphere's dynamics based on observations and the known physical explanations of the causes of circulations. Even though his work was criticized considering inaccurate and without scientific value, his many significant contributions to weather forecasting optimistically promoted the practical utilization of meteorology. Soon, meteorologists begun to map data from various observing stations such as temperature and humidity, onto weather charts. Availability of weather maps enabled scientists to detect and study various weather phenomena and compare the current meteorological situation to past ones – ultimately leading to more forecasts being produced. Because it became evident that the natural knowledge was inadequate, there was a need to further understand the dynamics of the atmosphere. For example, instruments were needed to measure properties of the atmosphere (e.g. moisture, speed, temperature) and how they vary with time. A great progress in monitoring weather was made in the 1920s after the invention of the radiosonde, and later a great achievement was the invention of the telegraphs in the mid-nineteenth century allowed routine transmission of weather observations between observers and compilers.

23.2.3 Modern Weather Forecast

Before the World War I, the weather forecast was mainly based on personal experience, intuition, and the subjective knowledge of the person responsible for issuing the weather forecast information. Thus, predictions were never done by employment of mathematical-numerical modelling of the atmospheric processes, while the further behaviour of the atmosphere was estimated by a set of empirical rules, transferred from one forecaster to another.

In the late nineteenth century, many scientists worldwide realized that the atmospheric processes could be simulated utilizing the fundamental physical principles. Knowing the basic initial state of the atmosphere and applying the laws of physics with integration of mathematical set of equations will help to obtain the further state of the atmosphere. This fundamental weather forecasting concept was altered dramatically in the 1920s by Norwegian scientist Vilhelm Bjerknes (see Fig. 23.3). He was convinced that the prediction of the future state of the atmosphere and its evolution could be formulated mathematically using the main velocity (vector) variables and four scalar quantities (e.g. temperature, pressure, density, and humidity) each being function of space and time. In order to promote and develop the atmospheric modelling concept and its physical structure, a famous Bergen School of Dynamic Meteorology was established under Bjerknes' leadership. His scientific work was

Fig. 23.3 Vilhelm Bjerknes (1862–1951)

Fig. 23.4 Weather Forecasting Factory by Stephen Conlin, 1986. (Credit: Hendrik Hoffmann, University College Dublin)

focused on determination of physically basis for weather forecasting, considering that two important elements should be satisfied:

- A good knowledge of the initial status of the atmosphere (initial conditions)
- A good knowledge of the fundamental principles according to which atmosphere changes its state from one to another

Bjerknes' concept was later developed by Lewis Fry Richardson, who is considered as the father of numerical weather prediction (NWP). In 1922, Richardson published a remarkable book titled "Weather Prediction by Numerical Process". This pioneering work focused on the requirements for a successful numerical integration (the method and computer performances), resembling a forecast factory (Wiston and Mphale 2018). Several artists have created images of his idea on the forecast factory. Conlin's image (see Fig. 23.4) clearly depicts a large building, with a central chamber. On the wall of this chamber is a gridded map with half the globe, including red grids which denote pressure and white cells (wind). According to Richardson, the scheme was complicated because of the complex nature of the atmosphere, required for accurate and detailed implementation of the Bjerknes' concept. He concluded that all computations had to be done manually, estimating that it would take about 64,0000 human computers to perform the calculation in time with the weather happening. For example, his work took several months of calculations of pressure changes to produce a 6-hr forecast. His *failure* was later shown to be due to lack of understanding of some atmospheric processes at the time and the initial conditions. To date meteorological observations are made all over the world and are used to compute the best estimate of the system's initial conditions. Although it was a spectacular failure in the end, Richardson devised many of the modern principles of numerical weather forecast. Increasing computing power and efficient numerical methods as well as more sophisticated physical parameterizations has led to a huge improvement of weather forecasts.

23.2.4 Early Numerical Weather Prediction (NWP)

Numerical weather prediction (NWP) represents an advanced method of weather prediction by use of computational techniques through time integration of the fundamental equations or computer programs. The process describes a set of partial differential equations (PDEs) and other formulations describing the dynamic and thermodynamic processes in the Earth's atmosphere, comprising of equations, numerical approximations, parameterizations, domain settings, as well as initial and boundary conditions. The physics laws are programmed to model the atmosphere, written in a format that the computer can understand. The earliest work on NWP can be traced back to the contribution of Cleveland Abbe, an American meteorologist in the 1890s, who recognized that "meteorology is essentially the application of hydrodynamics and thermodynamics to the atmosphere".

His goal was to make meteorology an exact science and a true physics of the atmosphere utilizing a graphical, analytical, or numerical method to solve equations. The idea was later taken up by Bjerknes using diagnostic and prognostic methods and variables to predict weather, despite some limitation based from the inadequate weather information in some areas.

He suggested that it would be possible to forecast the weather by solving a system of nonlinear PDEs, although the mathematical models he proposed were far too complex to be solved analytically. However, modern concept of numerical weather forecasting was initiated by Richardson during World War I, developing the concept of both Abbe and Bjerknes and using his own theories and reasoning. Richardson presented a set of equations describing the physical processes governing atmospheric behaviour, together with a method for their approximate solution, arguing that it should be possible to proceed from an initial to a final state of the atmosphere by a purely mathematical process (Lynch 2014). Although Richardson's example was a spectacular failure, his method revolutionized modern numerical weather

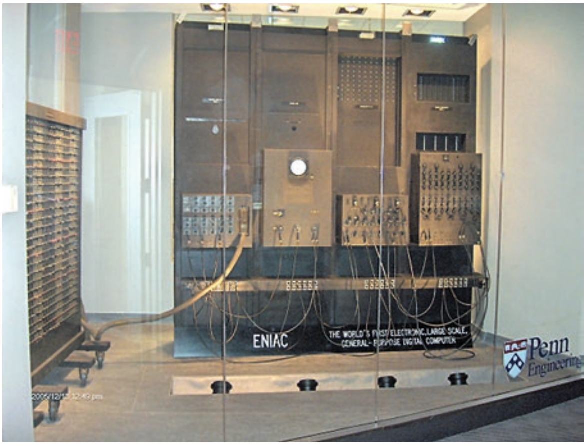

Fig. 23.5 The earliest electronic computer ENIAC. (Credit: Sakisbal [CC BY-SA 4.0 (https://creativecommos.org/licenses/by-sa/4.0)])

forecasting. The concept was developed further in the 1940s by Charney and colleagues, providing a theoretical basis to overcome problems faced by Richardson. The simplified equations they proposed lead to the construction of an electronic computer ENIAC (see Fig. 23.5), which became a milestone in NWP. The first computer generated prediction of the flow in the middle atmosphere was developed in 1949 by Charney, Fjortoft, and von Neumann. They used a simplified form of atmospheric dynamics based on solving the barotropic vorticity equation over a single layer of the atmosphere, by computing the geopotential height of the atmosphere's 500 hPa pressure surface. In September 1954, Carl-Gustav Rossby assembled an international group of meteorologists in Stockholm and produced the first operational forecast (i.e. routine predictions for practical use) based on the barotropic equation. Operational numerical weather prediction in the United States began in 1955 under a joint project by the US Air Force, Navy, and Weather Bureau. The Japanese Meteorological Agency became the third organization to initiate operational NWP in 1959. The first real-time forecasts made by Australia's Bureau of Meteorology in 1969 for portions of the SH were also based on the single layer barotropic model. Later models used more complete equations for atmospheric dynamics and thermodynamics. In 1959, Karl-Heinz Hinkelmann produced the first reasonable primitive equation forecast, 37 years after Richardson's failed attempt. In 1966, West Germany and the United States began producing operational forecasts based on primitive-equation models, followed by the United Kingdom in 1972 and Australia in 1977. Later additions to primitive equation models allowed additional insight into different weather phenomena including solar radiation in 1967; moisture effects, latent heat, and convection were incorporated in 1971. Three years later, the first global forecast model was introduced. Sea ice began to be initialized in forecast models in 1971.

23.3 Numerical Weather Prediction Concept

Numerical weather and climate prediction models (Palmer and Hagedorn 2006) comprise of fundamental laws and parameterized physical and chemical components of the atmosphere. The state of the atmosphere is described at a series of “grid points” by a set of variables such temperature, pressure, velocity, and humidity. The laws are expressed as mathematical equations, averaged over time, and grid volumes – describing the evolution of such variables. They are solved by replacing time derivatives by finite differences and spatially either finite difference schemes or spectral methods (i.e. state of the body as a function of time). They are converted into a computer program, defining among other things, possible integrations between the variables with other formulations, and integrated forward in discrete time steps (i.e. making them predictive) to simulate changes in the atmosphere. In this context, the model is a computer program that produce meteorological information at given locations. All numerical models are based on the same set of governing laws used to predict the physics and dynamics of the atmosphere. Mathematical formulation of atmospheric models used in weather forecasting is based on equations mechanics of compressible fluid, originating from three fundamental laws: momentum, mass conservation, and thermodynamic equation. Various physical quantities that characterize the state of the atmosphere are assumed to have unique values at each point in the atmospheric continuum. Moreover, these field variables (e.g. pressure, density, temperature, speed) and their derivatives are assumed to be continuous functions of space and time. These include ideal gas law given by:

$$\mathrm{p} = \rho RT \tag{23.1}$$

where p is pressure, ρ is density, R is gas constant (287 $\mathrm{Jkg}^{-1}\ \mathrm{k}^{-1}$) – a universal quantity that does not change with time, and T is temperature. Hydrostatic equation given by:

$$\frac{d\phi}{\mathrm{dp}} = -\frac{\mathrm{RT}}{\mathrm{p}} \tag{23.2}$$

where Φ is the geopotential at a given height/altitude, T is temperature, and p and R are quantities, defined above. This means that the thickness of an atmospheric layer bounded by isobaric surface is proportional to the mean temperature. Pressure decreases more rapidly with height in a cold layer than a warm layer. Rate of change of wind or momentum equation given by:

$$\frac{\mathrm{d}\mathbf{V}}{\mathrm{dt}} = -\frac{1}{\rho}\nabla \mathrm{p} - 2\Omega x\mathbf{V} - \mathrm{g}\mathbf{k} + \nu\nabla^2\mathbf{V} \tag{23.3}$$

where 2 Ω × V is the Coriolis force (Ω is the angular velocity of planetary rotation and V = **i**u + **j**v + **k**w is the velocity vector); the Coriolis parameter is given by f = 2Ωsinφ. ∇p is the pressure gradient, g is gravitational field strength, **k** is the vertical unit vector, and $\nu\nabla^2 V$ is the fluid viscosity. Thermodynamic equation, given by:

$$c_v \rho \frac{dT}{dt} + p\nabla \cdot V = Q_h + Q_d \tag{23.4}$$

where Q_h and Q_d represent adiabatic and diabatic heat fluxes of air. The heat could be added to a unit air mass due to external sources (e.g. solar and thermal radiation, turbulent heat exchange or phase transformation of atmospheric moisture). Conservation of water mass given by:

$$\frac{1}{\rho}\frac{d\rho}{dt} + \nabla \cdot V = 0 \tag{23.5}$$

where ∇ is the three-dimensional gradient operator and V is the 3D wind. Potential temperature given by:

$$\theta_s = T\left(\frac{p}{p_0}\right)^{-k} \tag{23.6}$$

where $k = \frac{R}{c_P}$, c_P is specific heat capacity at constant pressure and R is as defined-earlier. This relationship is referred to as Poisson's equation, where temperature θ_s is called the Potential temperature. θ is simply the temperature that a parcel of dry air at pressure p and temperature T would have if it were expanded or compressed adiabatically to a standard pressure p_0 (1 atm = 1.01325 × 105 Pa). For dry air, $\kappa = 0.286$. Numerical weather prediction has been viable since the 1960s. Before the computer era, principles of theoretical physics played little role in practical forecasting. Although much of the underlying physics was known, its application to predict atmospheric conditions was impractical. Key development in weather forecasting was made in the immediate post-war period after the introduction of beginning of computer technology to solve the complex equations and put the theory of dynamic meteorology into practice. For example, Bjerknes theorized that the atmosphere must obey the basic laws of physics. He stated (1904) that "subsequent atmospheric states develop from the preceding ones according to the physical law". By stating the laws as mathematical equations, real observations from the atmosphere could be used to generate a mathematical model to simulate the atmosphere. Qualitative data (current state of the atmosphere) is collected, from which scientific understanding of the atmospheric processes is used to infer how the atmosphere will evolve over time. Since some physical processes take place on scales smaller than the horizontal grid scale of the model domain, for these reasons they cannot be easily resolved in the models. In some cases, they are explicitly accounted for while in other cases may be neglected or calculated appropriately as corrections. This means effects of physical processes are included implicitly (or indirectly) when they cannot be included explicitly. The approximation of such unresolved processes is referred to as parameterization. Processes associated with this include radioactive transfer, microphysics (e.g. moist processes such as cloud and rain drop), turbulent mixing, and atmospheric emissions (e.g. aerosols and gases, air quality). There are many types of parameterization schemes used in different models. These are important aspects that strongly influence model forecasts and interact with each other in the atmosphere. Parameterization needs to account for computational costs (e.g. model running time, resolution used and/or any other resource specification), especially when using or increasing the complexity of the choice. There must be a "balance" between the type of scheme chosen and the computational cost. Data is interpolated onto a global grid comprising of several grid points. The planet is divided into a 3D grid. The basic equations are applied and results evaluated to calculate variables such as wind field, heat transfer, radiation, temperature, pressure, etc. within each grid. However, there could be cases where data is insufficient or severe, measurements not covering the entire globe and/or not at set points. On the other hand, because observations cannot be used directly to start the model integration, they are modified in a dynamically consistent way to obtain a suitable dataset. Input data need to be interpolated, smoothed, and filtered a process usually referred to as data assimilation. Real observation data is compared with predicted conditions to give the best possible estimate of the atmosphere. Many studies have devoted to improving numerical modelling through more advanced numerical methods, better representation of atmospheric dynamics, and improved parameterization schemes. However, these developments can be inhibited by inadequate observations, limited understanding of the atmospheric physical processes, and the chaotic nature of the atmospheric flow. Therefore, uncertainties will always exist in both initial conditions and/or numerical prediction results. Many models are developed to improve precision and to accommodate more complex processes.

23.3.1 Uncertainties in Weather Forecasts

There are several uncertainties why weather forecasts sometimes go bad or wrong. First, it comes from the initial conditions and the models (tools) themselves. Forecasting tools and other methods used can vary with the time period of forecast. For example, the chaotic nature of the atmosphere, computational power required, errors involved in measurements, and incomplete understanding of the atmosphere show that forecasts become less accurate as the difference between the current and the time for which the forecast is being made increases. It is also worth noting that measured or observed data (including initial conditions) and atmospheric equations are used to forecast the future status of the atmosphere. Therefore, measurements might not be sufficiently precise or detailed, biasness in the choice of weather stations, and computer programs not storing data to infinite precision. Uncertainty due to parameterization of sub-grid scale processes also plays a crucial role in prediction quality. Furthermore, models include some equations accounting for the effects of small-scale processes that cannot be explicitly represented as their resolutions may not be high enough. This can directly or indirectly affect accuracy in the predictions. Consequently, the further you go into future, the model's precision skill is lost. Model prediction accuracy is limited within a few days. For example, Edward Lorenz claimed that even with a perfect model stating with initial conditions, weather forecasting is limited to about 2 weeks only. He argues that no matter how good the observational network or how good the forecasting procedures, there is almost certainly an impossible limit as to how far into the future one can forecast as the small errors will always exist and can grow into large errors within a certain amount of time. Some of the factors influencing numerical weather prediction and thus on accuracy and why do weather forecasts sometimes go bad from (a–b) are listed below:

(a) Limited knowledge of the current state of the atmosphere:

- Insufficient or lack of interesting data in between observing locations
- Instrument sensitivity and errors
- Lack of observation of important quantities
- Errors in data assimilation (imperfect statistical and numerical forecast methods)

The overall impact of this can lead to limitations in observations. For example, satellites, radars, and radiosondes might not provide all the information missed by surface observations. Similarly, data inconsistencies between different tools, incomplete understanding of the complexities, and interactions between atmospheric physics and chemistry can be another cause of errors in the forecasting.

(b) Imperfect computer model:

- Need more powerful and faster computers for more grid points
- Need of better understanding of physical processes going into the model grid
- Limited resolution (processes represented must be truncated spatially, temporally and physically)
- Systematic and/or random errors

Computer forecasts can be deficient in that they neglect small-scale effects and/or approximate complicated physical processes, for example, processes occurring at scales smaller than horizontal model grid sizes. There is also a significant source of random errors associated with parameterization.

(c) Chaos:

- Sensitivity dependence to initial conditions (the atmosphere could react very differently to slightly different initial conditions) – the "butterfly effect".
- Small differences in initial conditions can evolve into large changes/amplify rapidly (e.g. forecasts skill is lost with increasing lead time and is case specific).

The word "chaos" was used to describe the model sensitivity to initial conditions – a property popularly known as the "butterfly effect" that the atmosphere exhibits a dynamical sensitivity such that predictions would lose skill at some point in the future. This was captured in the "chaos theory" pioneered by Lorenz in the 1960s. Golestani and Gras also note that chaotic behaviours are strongly dependent on initial conditions and those small changes in initial conditions can possibly lead to immense change in subsequent time steps and particularly difficult to predict. Some of the problems encountered in weather forecasting could also arise from the fact that we don't fully understand everything that happens in the atmosphere. A wide variety of factors influence the weather in many ways. The fact that a number of relevant processes occur at scales smaller than model grids would possibly introduce uncertainties into the model as we don't always have a good understanding of the behaviour of such processes, particularly their response to feedbacks. However, it is worth noting that in the real world, a wide variety of forces are in play at the same time. Therefore, it can be difficult (and sometimes impossible) to identify dominant factors in the system. Although observations, theories, models, and other tools continue to improve, the Earth-atmosphere is an immensely complex system.

23.3.2 Weather Forecast Practice

The task of determining the future state of the atmosphere is called weather forecasting. To develop weather conditions, it is necessary to analyse them. The procedure for analysis involves assessment of current atmospheric conditions, which include collection, transmission, data processing, and application (Sene 2010; Lackmann 2012). The components of the modern system for forecasting the weather include:

- Data collection
- Data assimilation
- Data reanalysis
- Numerical weather forecasting
- Products of the model outputs (post processing)
- Presentation of the forecast for end users

In the global scale, the World Meteorological Organization (WMO) is responsible for collecting, drawing, and dissemination of weather data. Information once collected is distributed to the three regional meteorological centres in Washington DC, Moscow, and Melbourne, Australia. The process of providing weather forecasts and warnings is conducted in three stages. First, data are collected and analysed on a global scale. Then, different techniques are used to determine the future state of the atmosphere, a process called weather forecasting. Finally, the weather forecast outputs are delivered to the public and other end users.

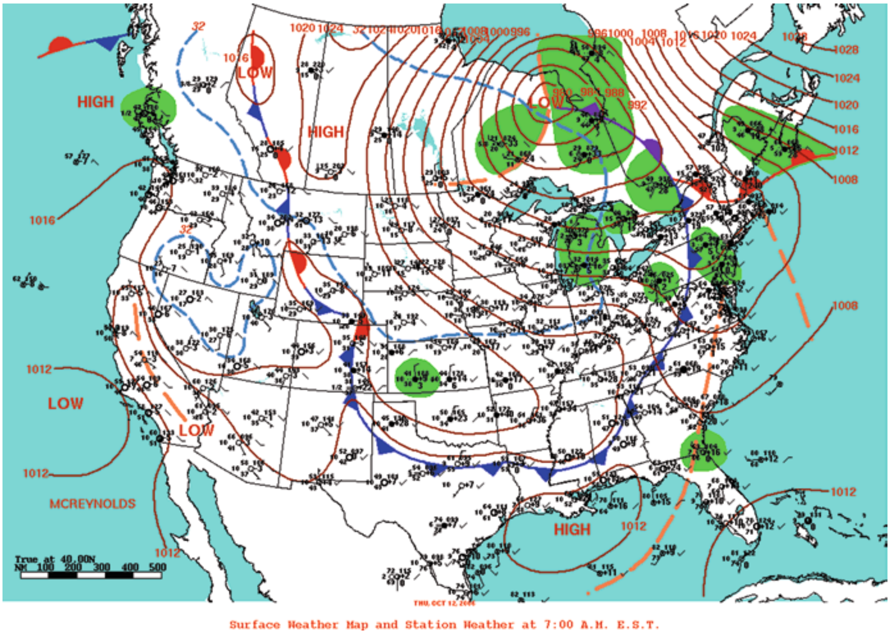

Fig. 23.6 Surface weather chart. (Credit: Hydrological Prediction Center of NOAA. Surface chart the 12th of October 2006 showing the cold air front intrusion over the southern Great Lakes, causing a Snow Storm with heavy snowfall of about 60 cm in a narrow area surrounding Buffalo, NY)

23.3.3 Weather Charts

Normally, a large quantum of the available weather data once collected is shown in synoptic (observations made at the same time) *weather chart*. Weather chart shows the status of the atmosphere, including information about temperature, humidity, pressure, direction, and speed of wind and other meteorological elements and phenomena. Figure 23.6 shows an example of the weather chart.

23.4 Weather Forecast Methods

Weather forecasts today are prepared by both objective and subjective methods.

23.4.1 Objective Methods

An objective forecast is one that is made without recourse to the personal judgement of the forecaster. Strictly speaking, if two forecasters were given copies of one manual describing a forecast method and placed in separate rooms with current data, they would make identical forecasts. Its primary purpose is to bring forecasters closer to the elusive goal of accuracy. In an objective method, meteorological parameters appear in formulas or relations that are theoretical or empirical, in graphical, tabular, or algebraic forms. These methods generally involve the use of numerical (i.e. physical/dynamical) and/or statistical models. These methods are objective in the sense that, for a procedure and set of relevant data, the forecasts produced do not depend on a meteorologist's judgement, although subjectivity is involved in the choice of a procedure and a set of data.

23.4.2 Subjective Methods

These are methods in which the formulation of the forecasts is based at least in part on the judgements of one or more meteorologists. Subjective methods are those in which the processes used to analyse the data have not been well specified. These methods are also called implicit, informal, intuitive methods.

They may be based on simple or complex processes; they may use objective data or subjective data as inputs. Weather forecast is said to be the goal of atmospheric research. It is also described as the most advanced area in meteorology. The nature of modern weather forecasting is not only highly complex but also highly quantitative. An attempt, therefore, is made to highlight the different methods used in modern weather forecasting. The various methods used in forecasting the weather are as follows:

1. Synoptic weather forecasts
2. Numerical methods
3. Statistical methods

Synoptic weather. Synoptic weather conditions are a principal method for preparing weather forecasts, which dates from the end of 1950, and it includes an analysis of synoptic weather maps, using several empirical rules. This is the first method for preparing weather forecasts. The meaning of the word "synoptic" means that the observation of different weather elements refers to a specific time of observation. Thus, a weather map that depicts atmospheric conditions at a given time is a synoptic chart to a meteorologist. In order to have an average view of the changing pattern of weather, a modern meteorological centre prepares a series of synoptic charts every day.

Numerical weather prediction. Modelling of the atmosphere is top technology in meteorology (Jacobson 1999). It is believed that one of the tasks of every modern state is to use products from these automated systems and, if possible, to have an operational control. Top in this area are those countries which can make models of the atmosphere and thus to establish a full insight of the events of the atmosphere, at least over its geographical area. All of them are based on one of the most successful approaches in addressing hydrodynamic equations for the atmosphere, which found practical application and verification in many known and recognized models

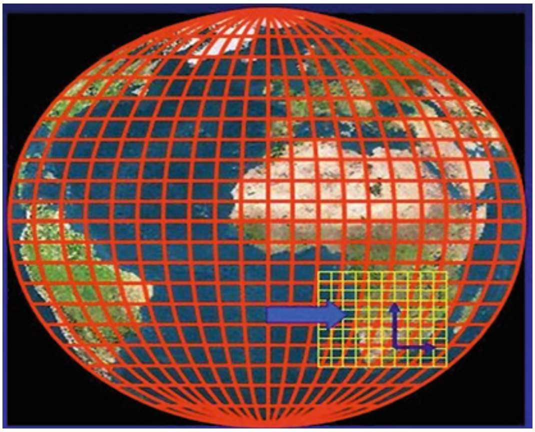

Fig. 23.7 Numerical weather-climate prediction model (Example of numerical grid-point model that covers 3D domain with grid of points, solve forecast equations at grid points)

of the atmosphere worldwide. Numerical weather conditions are widely used in modern weather forecasting. Since atmospheric processes obey a few known physical laws and principles, they include the law on maintenance of energy, momentum, and continuity. Ideally, these physical laws can be used to forecast the future state of the atmosphere, using certain initial and boundary conditions, depending on geographical area, elevation, vegetation, and the meteorological parameters and products derived from the global system of monitoring and the initial fields from global models. This is called "Numerical Weather Report" because of the mathematical way to represent all these important atmospheric processes, arising from the laws, and the numerical calculated using computer algorithms.

Numerical weather conditions use a number of highly refined computer grid point (see Fig. 23.7) or Lagrangian models that try to represent the behaviour and the future stage of the atmosphere.

Statistical methods to forecast the weather. Statistical methods use historical weather data to forecast future cases. They are often used in combination with numerical weather forecasts. A statistical approach, method of analogy, analyses past weather reports to find ones that have approximately the same duplicate conditions. Statistical methods are used along with the numerical weather prediction.

Statistical methods use the past records of weather data on the assumption that future will be a repetition of the past weather. The main purpose of studying the past weather data is to find out those aspects of the weather that are good indicators of the future events. After establishing these relationships, correct data can be safely used to predict the future conditions. Numerous attempts have been made to use statistical methods to identify the main drivers of changes in the atmosphere and to use them for making long-term forecasts. These techniques were made in two directions:

1. To determine the time periodicity in some occurrences or values of meteorological elements.

2. To find any "interactions" between certain factors in the atmosphere. In the statistical methods, the following sub-classification exists:

   - Method of symmetrical points
   - Method of singularity
   - Method of types
   - Method of analogue
   - Method of natural synoptic period
   - Combined physical-empirical methods

Method of symmetric points. German meteorologist Vikman (1931) proposed a method of forecast based on symmetrical points. A symmetric point is defined as a case of the pressure graph when the curve is repeated. The appearance of the inverse symmetry of the pressure curve is also frequent. Symmetry usually occurs when at high amplitudes, the peaks of the valley valleys are swirling and are inversely when the null points agree.

Method of singularity. In the literature, many papers can be found that support the thesis that some time types appear more correctly than could be expected on the coincidence of chance. Thus in meteorology the term "singularity" is defined, which is defined as a confidential time phenomenon that almost every year repeats in approximately the same time.

Method of types. In many countries, the elaboration of mid-term and long-term forecasts was based on the method of types. The German meteorologist Baur (1947) has the greatest contribution to the development of the method of types. The method of types is defined as "the mean distribution of sea level pressure over a time interval, when the positions of the stationary cyclones and anticyclones remain substantially unchanged".

Method of analogue: If two similar weather situations are found, their further development will be similar. Accordingly, if we find a map like the current weather map in the synoptic archive, the forecast for the next period can be given directly by reading the weather from the series of maps that follow the analogy map in the past. The method of analogy is a quick, inexpensive, and objective way of making the weather forecast, which includes climatological characteristics and orographic influences over the area under consideration. However, the problem is that it is not possible to find two identical weather charts.

Method of natural synoptic period. The first and most important works in the former Soviet Union in the field of long-term forecast of time were given by Multanowski (1915). The basis of its synoptic long-term prognosis is the idea of the existence of "natural synoptic periods". According to the ideas of Multanowski, a natural synoptic period can be defined as a time interval during which the thermal and barrier fields in the atmosphere are not significantly changed, which condition a certain orientation of the movement of the bar systems on the Earth's surface and maintain a geographical distribution of the signs of the bar field in the area of the natural synoptic region.

Combined physical-empirical methods. American meteorologist Namias (1947) proposed medium-term and long-term prognosis for the method of average atmospheric circulation. This method assumes that the general circulation index, which is the volume of zonal and meridian air masses, has its own repetition cycle. His method uses combined physical-empirical methods with four different cycles.

23.5 Weather Forecast Types Based on the Range of the Processes

To produce a weather forecast, we need to model the dynamics of the atmosphere and the physical processes that occur, such as the formation of clouds, and the other processes in the Earth system that influence the weather such as atmospheric composition, the marine environment, and land processes. Any forecast we produce is limited by the fact that the atmosphere is chaotic (we can never fully know the exact initial state of the atmosphere) and that our numerical models cannot perfectly represent the laws of physics governing the dynamic equations. We also must simplify our models of many processes which occur at very small scales, such as cloud formation. This means that all forecasts will have some uncertainty associated with them. There are four types of operational weather forecasting based on range of various atmospheric processes (Inness and Dorling 2010):

1. Very short-range weather forecast (nowcasting)
2. Short-range weather forecast
3. Medium-range weather forecast
4. Long-range weather forecast

23.5.1 Very Short-Range Weather Forecast

This advanced technique, often called nowcasting, or very short-range weather forecasting, can provide a forecast up to 2 hours in advance. The current weather is mapped, and then an estimate of its movement is used to forecast the weather a short period ahead (assuming the weather will move without significant changes). It takes time to gather and map weather observations, so a short forecast is needed to outline what the weather is currently. The forecast utilizes the latest radar, satellite, and observational data to quickly predict hazardous weather cases, such as storms, tornadoes, hailstorms, and other weather phenomena. In addition cloud-resolving model simulation (see Fig. 23.8) could be a useful tool in simulation of convective clouds and thunderstorms. The severe weather features relevant for very short-range forecasting are:

– Heavy precipitations episodes
– Individual convective events (squall lines, organized convection)
– Generalized freezing precipitations

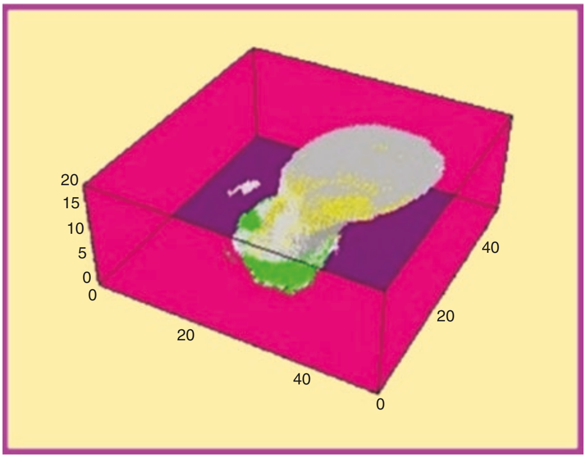

Fig. 23.8 A three-dimensional simulation of severe storm using cloud model (Numerical simulation of Skopje supercell storm as individual convective event on 6 Aug 2016 using a cloud resolving model)

– Snow storms and streamers – windstorms
– Fire danger
– Severe heat or cold weather
– Dust storms

23.5.2 Short-Range Weather Forecast

Short-range weather forecast is beyond 12 hours and up to 72 hours. Figure 23.9 shows an example of a short-range synoptic scale forecast of precipitation patterns of the tropical cyclone Irma on September 8, 2017. The simplest techniques for short-term weather conditions are called consistent or permanent, predicting that future weather will be the same as the current conditions.

The severe weather features relevant for short-range forecasting are:

– Heavy precipitations episodes
– Areas of convective events (probability of occurrence)
– Snow storms or streamers
– Windstorms
– Heat wave or cold wave (start, continuation or end)
– Fire danger

23.5.3 Medium-Range Weather Forecast

Medium range represents the weather forecast for several days to 2 weeks in advance (Woods 2006). Figure 23.10 shows an example of a medium-range weather forecast chart. The ability to represent location and time of the time cases is reduced by increasing the length of the forecast. In the current situation, the medium-range

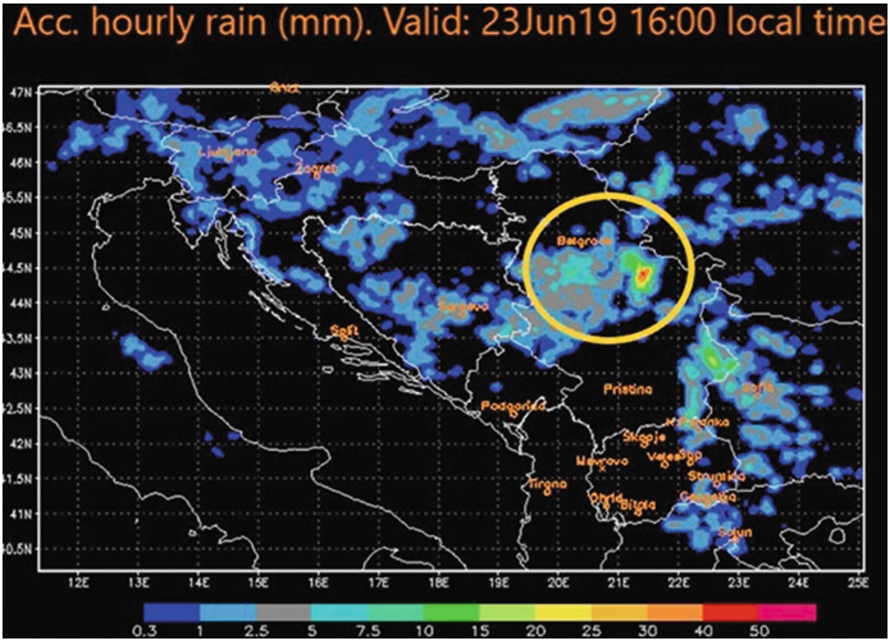

Fig. 23.9 A short-range forecast of heavy rain (WRF-ARW Forecast of hourly rain during severe flooding occurred in Belgrade, Serbia, on June 23, 2019)

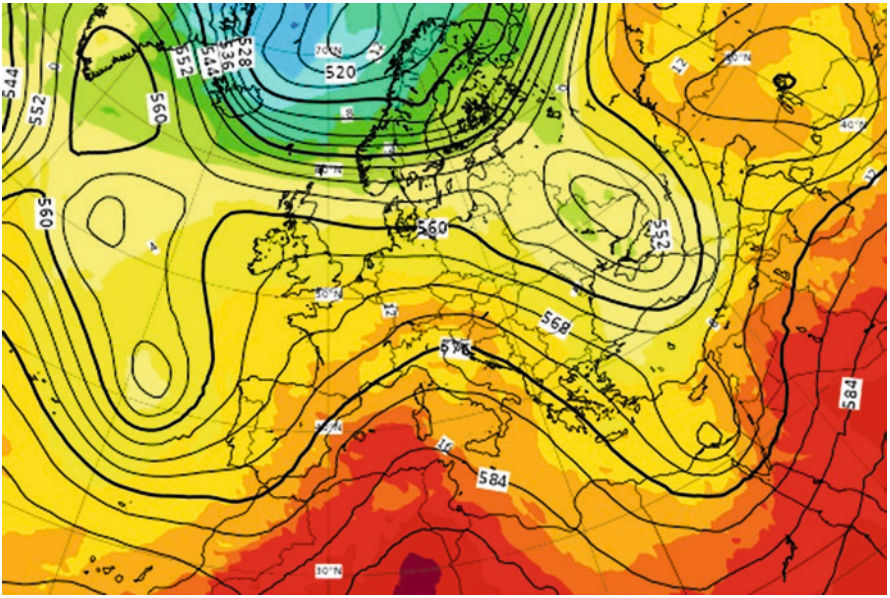

Fig. 23.10 Medium-range weather forecast. (Credit: ECMWF)

forecasts are based primarily on global systems for numerical estimates. On average, these systems produce skilful forecasts of more than 1 week in advance, although their performance is changing the season in the region. The severe weather features relevant for medium-range weather forecasting are:

(a) Heavy precipitations episodes (early warning and probability of occurrence)
(b) Areas of convective events (early warning and probability of occurrence)
(c) Snow storms (early warning and probability of occurrence)
(d) Windstorms (early warning and probability of occurrence)
(e) Heat wave or cold wave (start, continuation, or end)

Forecasters use the basic methods in order to increase the skill of medium-range forecasts for a day or more and get a picture of the potential skill of the forecast before it is verified. General forecasts are forecasts from the complex system of weather prognostic systems with slightly different initial conditions and/or some different versions of atmospheric models. Differences between individual forecasts generally show probable prognostic skill and reliability. Using the general prognosis in a significant increase is common for all the revolutionary short-term processes, and it is predicted to change. In the last three decades, sophisticated medium-range forecasts extend roughly over 1 day after decade. The main winter storms can now be predicted one or more weeks in advance, allowing the maintenance of the road by the staff and enough time to prepare the managers responsible for such emergency weather situations.

23.5.4 Extended-Range Weather Forecast

This type of forecast is defined in the Manual on the GDPFS as a description of weather parameters beyond 10 days and up to 30 days (monthly weather forecast) usually averaged and expressed as a departure from climate values for that period. It is often made as an extension of ensemble prediction systems for medium-range weather forecasting, with a lower spatial resolution as the time range is extended to 1 month. Deterministic forecasting is not possible for extended-range weather forecasting because the uncertainty is such that only probable departures from climate values can be given along with a rough indication on the probability of realization. The occurrence of extreme events cannot be accurately forecast by such systems. Extended-range weather forecasting will take advantage of using the different EPS systems available in the world. Figure 23.11 shows an example of the extended range weather forecast provided by ECMWF, using a combination of different output parameters. The time range 10 to 30 days is probably still short enough that the atmosphere retains some memory of its initial state and it may be long enough that the ocean variability has an impact on the atmospheric circulation.

Therefore, extended-range weather forecasting can be built as a combination of the medium-range ensemble prediction systems and seasonal forecasting systems and contain features of both systems, like coupled ocean-atmosphere integrations.

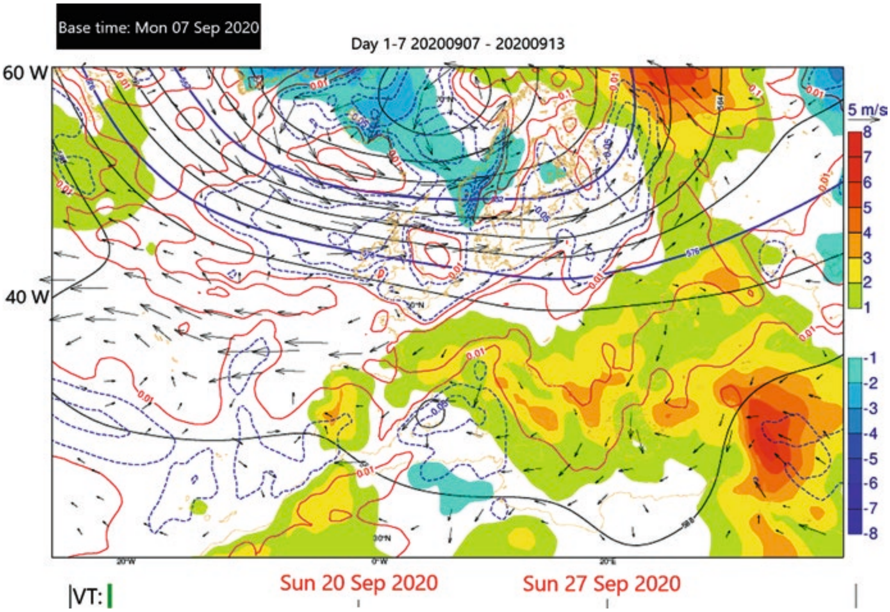

Fig. 23.11 Extended-range weather forecast using multiparameter outlook. (Credit: ECMWF)

The severe weather features relevant for extended-range weather forecasting are:

(a) Higher or lower than normal precipitation episodes
(b) Higher or lower than normal temperatures: can be associated with, e.g. ENSO events, cold waves, heat waves, and droughts

23.5.5 Monthly Weather Outlooks

Monthly outlook is a description of averaged weather parameters expressed as a departure from climate values for that month. Monthly outlook can be obtained by running ensemble prediction systems at a larger scale than for medium-range forecasting – for computer resources reasons, nevertheless allowing for usable trends up to a month in advance. The analysis of multi-model ensembles is useful in long-range forecasting. The severe weather features relevant for monthly outlooks are higher or lower than normal temperature or precipitation episodes (start, continuation, or end).

23.5.6 Long-Range Weather Forecast

Long-range or global weather forecast represents a special type of forecast which usually a global range for an extended period of 1 year and more. Figure 23.12 shows a mean sea level pressure (hPa) as a product of the Global Forecast System

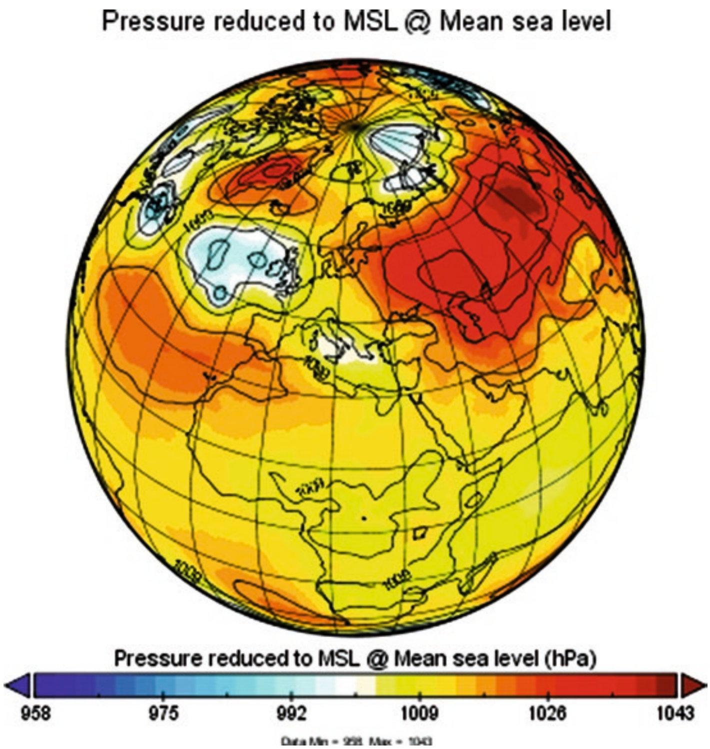

Fig. 23.12 Global weather forecast

Model (GFS). Continuous time forecasting is an area that is firmly based on statistical averages derived from past cases; weather is often referred to as climate data. Weekly, monthly, and seasonal weather outlook prepared by the National Weather Service weather forecasts does not represent the general sense. They indicate only whether the region will hit near-normal precipitation and temperatures or not. Ability and skill of forecasting rely on knowledge, experience, and analytics of a forecaster.

23.5.7 Climate Forecast

Climate forecasting covers forecasting ranges beyond 2 years:

- Climate variability prediction is a description of the expected climate parameters associated with the variation of interannual decadal and multi-decadal climate anomalies.
- Climate prediction is a description of expected future climate including the effects of both natural and human influences. Climate forecasting can be achieved by using coupled models which encompass atmosphere and ocean physical pro-

cesses and their interaction (see Warner 2011). However, as the time range is higher than 90 days and can reach several years if not centuries, slow-varying phenomena must be considered to adequately represent climatic trends, with, for instance, model representations of polar ice fields or vegetation change. Thus, climate models can be built as coupled models with many different "bricks" interacting each other: atmosphere, ocean, ice fields, aerosols, vegetation, atmospheric chemistry, and other. Then, relying on economic studies, different scenarios of emission of greenhouse gases are used as an input in climate models so that the climate trends can be tentatively drawn following these scenarios.

23.6 Ensemble Forecast Method

Ensemble forecast represents a technique based on running several forecast models for the same weather situation or different versions of a single model, each beginning with slightly different initial condition uncertainty as it is shown on Fig. 23.13. Then the forecast output results are superimposed on the same graph, where the most common outcome is probably the most likely one. It is no doubt that the ensemble approach reflects the errors in the measurements, reducing the forecast uncertainty and thus significantly improves the forecast accuracy.

23.6.1 Deterministic Vs Probabilistic Forecast

Forecasts are either deterministic or probabilistic (Fig. 23.14). Both techniques try to predict events, but information on the uncertainty of the prediction is only present in the probabilistic forecast.

Deterministic forecasts of an event of a specific magnitude at a specific time and place. As we go from the short to the long term, the nature of the forecasts (or whatever we might call them) also changes. Short-term forecasts are deterministic.

Probabilistic forecasting is a technique for weather forecasting that relies on different methods to establish an event occurrence probability (Palmer 2017). This

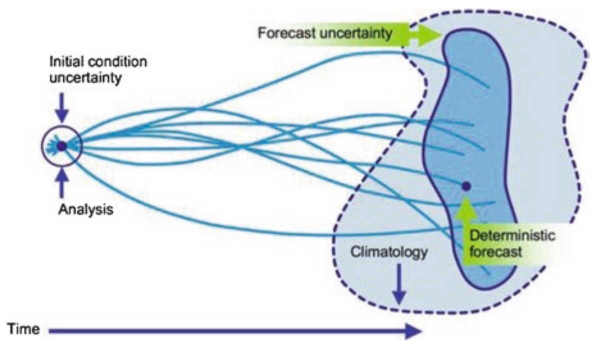

Fig. 23.13 Ensemble forecasting method

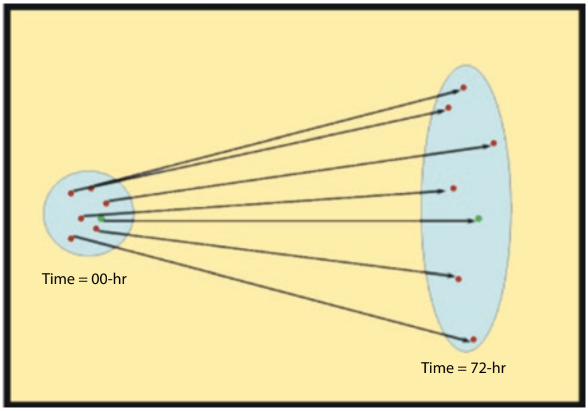

Fig. 23.14 Deterministic vs probabilistic forecast

differs substantially from giving a definite information on the occurrence (or not) of the same event, technique used in deterministic forecasting.

Traditionally numerical weather prediction has advanced progressively by improving single, “deterministic” forecasts with increasing model accuracy and decreasing initial condition errors. However, the meteorological atmosphere is a chaotic system on time scales of a few days and weeks, depending on the spatial scales of interest. (The climatic system is also chaotic, but on much longer timescales.) Also the behaviour of our numerical simulations of the atmosphere would continue to be affected by the problems typical of model simulations of chaotic dynamical systems even if we could have perfect initial conditions, write perfectly accurate evolution equations, and solve them with perfect numerical schemes, just because of the limited number of significant digits used by any computer (Lorenz, 1963). Looking at the problem from a slightly more fundamental point of view, a forecast explicitly cast in probability terms is better not only because it provides the user with an estimate of the error “of the day” but because it is more “truthful”. So, a probability forecast conveys a message which explicitly reminds the user that there is always a forecast uncertainty which should be considered, computed, and considered when making any practical use of the forecast. In fact, even “deterministic” forecasts are probability forecasts in disguise, since an error bar (even if only an average error bar) can and should always be associated with it. That error bar implies a probability distribution of predicted future states around a central value.

23.7 Numerical Weather Prediction Structure

Numerical weather models (NWM) operate in three main phases:

1. Analysis (pre-processing)
2. Prediction (processing)
3. Visualization, verification (post-processing)

The analysis phase: A gridded, three-dimensional analysis is produced with a previous forecast and observations. The process by which the above are combined is called *data assimilation*.

The forecast a quantity F (e.g. temp, pressure, humidity, wind) at point x is given as:

$$\mathrm{F}(\text{at future time}) = \mathrm{F}(\text{present}) + \text{Change in}(\mathrm{F})\text{between present and future time}$$

or written in form of equation:

$$\mathrm{F}(\mathrm{x}, \mathrm{t} + \Delta t) = \mathrm{F}(\mathrm{x}, \mathrm{t}) + \frac{\partial F}{\partial t}(\mathrm{x}, \mathrm{t})\Delta t \tag{23.7}$$

where the first term on the r.h.s. of the above equation represents "initialization" or "initial conditions" (from observations) and the second one is a partial differential equation describing rate of change of F.

23.7.1 Data Assimilation

Data assimilation is a mathematical method which combines theory (in terms of a numerical model) with observations (Park and Xu 2009). There are many uncertainties in the initial conditions for a numerical forecast model. Therefore, it is a goal to interpolate sparse observation data using (e.g. physical) knowledge of the system being observed to run numerical model based on observed data. Depending on the goal, different approaches may be utilized. Data assimilation differs from other types of machine learning, image analysis, and statistical methods in that it uses a dynamical model of the system being analysed. Data assimilation initially developed in the field of numerical weather prediction. Numerical weather prediction models are equations describing the dynamical behaviour of the atmosphere, written in linearized form and the appropriate code into a computer program. In order to use these models to make forecasts, initial conditions are needed for the model that closely resembles the current state of the atmosphere. Simply inserting measurements into the numerical model grid points did not provide a satisfactory solution. Real-world measurements contain errors both due to the quality of the instrument, reliable data, and how accurately the position of the measurement is known. These errors can cause instabilities in the models that eliminate any level of skill in a forecast. Thus, more sophisticated methods were needed in order to initialize a model using all available data while making sure to maintain stability in the numerical model. Such data typically includes the measurements as well as a previous forecast valid at the same time the measurements are made. If applied iteratively, this process begins to accumulate information from past observations into all subsequent forecasts. Gridded atmospheric analyses are produced by combining the following:

1. A previous forecast
2. Forecast uncertainty
3. Observations
4. Observation uncertainty

The resulting analysis is the most likely state of the atmosphere based on the given information. There are different modern methods of data assimilation:

- Three-dimensional variation data assimilation (3DVAR)
- Four-dimensional variation data assimilation (4DVAR
- The ensembles Kalman filter (EnKF)

23.7.2 Reanalysis

Reanalysis represents a systematic approach to produce datasets for climate monitoring and research. Reanalysis are created via an unchanging data assimilation scheme and model(s) which incorporate all available observations every 6–12 hours over the period being analysed. This unchanging framework provides a dynamically consistent estimate of the climate state at each time step. The one component of this framework which does vary is the sources of the raw input data. This is unavoidable due to the ever-changing observational network which includes, but is not limited to, radiosonde, satellite, buoy, aircraft, and ship reports. Currently, approximately seven to nine million observations are ingested at each time step.

ERA-Interim represents a third-generation reanalysis. The successor to ERA-Interim, ERA5, is available back to 1979 as of January 2019. It provides hourly estimates of atmospheric variables, a horizontal resolution of 31 km and 137 vertical levels from the surface to 0.01 hPa. It utilizes the best available observation data from satellites and in situ stations, which are assimilated and processed using ECMWF's Integrated Forecast System (IFS). The dataset provides all essential atmospheric meteorological parameters: air temperature, pressure, and wind at different altitudes, along with surface parameters like rainfall and soil moisture content and sea parameters like sea surface temperature and wave height.

Prediction (processing): The processing phase of NWP involves calculating the future state of the atmosphere (starting point = the analysis) under the following governing equations:

1. Conservation of momentum
2. Conservation of mass
3. Conservation of energy

Visualization and verification (post-processing): The Model Output Statistics (MOS) is a post-processing technique that correlates relationships between a model forecast and the observational data based on numerous forecasts. Depending on the target audience, it uses different format of output data, visualization tools and animation techniques. The forecast verification is a standard technique, applied for verifying the forecast quality and accuracy (Jolliffe and Stephenson 2003, 2011). This process is based on using standard mathematical statistics formula as mean error (ME), the mean systematic error (BIAS), mean absolute error (MAE), root mean square error (RMSE), correlation coefficient (CC), and other useful methods of comparing the forecast skills with observations.

23.7.3 Application of New Technologies for Weather Forecasting

Many technical advantages are made to improve the accuracy of the forecast. Automated systems for monitoring the ground floor are now used in places that are currently outside of the observation network. Microcomputer interactive systems allow forecasters to show, to manipulate, and to quickly assimilate large quantum and diversity of available data. Advanced radar networks such as NEXRAD (Last Generation Time Radar) are accompanied by Doppler radar, used for detection and positioning of weather phenomena in small proportions, such as tornadoes and storms. Weather forecasting is firmly based on information supplied by polar weather satellites and geostationary satellites. Their primary importance is to help in filling the missing data from observations, especially concerning the oceans. Weather satellites can generate several types of images, including visible and infrared images of water vapour. Satellite infrared images (images obtained from the radiation emitted by the object rather than reflected) help in determining the rainy regions within the cyclone. Future satellites will be capable of directly or indirectly detect the wind speed, humidity and temperatures at different heights.

23.7.4 Observations and Forecasts

For many years, meteorologists were aware of the strong correlation between cyclone disturbances at the surface and seasonal fluctuations in wave flow on western winds high above the Earth. Often, when upper level air flow creates waves with high amplitude and base flow from north to south, cold air moves south, and, in such conditions, cyclonic activity dominates the weather. On the other hand, when the flow is located west to east, moderate temperatures and several cyclone disruptions are experienced south of the jet stream. Although the effects of flow at a higher level are well documented, some unpredictable behaviour of the flow of the pitch holding long-maturity forecast is beyond the achievements of forecaster. The National Weather Service (NWS) provides weather, hydrologic, and climate forecasts and warnings. NWS is responsible for collecting and delivering information related to weather. Within the service centre, a forecast is established. Among the important services provided by the service is forecast and warnings for adverse weather conditions including storms, floods, hurricanes, monsoons, tsunami, tornadoes, winter inconvenient time, heat waves, and extreme warming. NWS data and products form a national information data base and infrastructure which can be used by other governmental agencies, the private sector, the public, and the global community.

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Chapter 24
Climate and Climate Change

24.1 Definition of Climate, Climate Elements, and Factors

The climate is a notion of average time for a region and for a longer period. While short-term weather describes the state of the atmosphere, the climate is a long-term pattern of weather in one area (Ahrens and Henson 2016). The climate has many more items from the average condition of the atmosphere, because the full description of the environment should also include changes and extremes in the atmosphere to accurately describe the overall character of the area. The important elements in the descriptions of the climate are the regime of temperature and the precipitation regime, because they have the greatest impact on people and their activities have effects on the distribution of vegetation and soil development. There are various factors that affect climate worldwide, among which the most important are:

- Distance from the sea
- Ocean currents
- Direction of prevailing winds
- Topography of the terrain (orography)
- Geographic latitude (i.e. the distance from the equator)
- Phenomenon "El Niño-Southern Oscillation"
- Human activity

Many scientific papers and publications (e.g. Oliver 2005; Allaby 2007; Barry and Hall-McKim 2014) are dedicated to the study of Earth's complex climate system, including climate change as one of the most significant global challenge of modern human life.

V. Spiridonov, M. Ćurić, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_24

24.2 Climate System

The key to understanding climate change is knowledge of the basic characteristics of the climate system and its way of operating. In a planetary scale, global climate depends on the energy that the Earth receives from the Sun (Allaby 2007; Barry and Hall-McKim 2014). Global climate changes affect the energy within the climate system. The physical components of the global climate system include the atmosphere, oceans, ice cap (cryosphere), living organisms (biosphere), and ground, sediments, and rocks (geosphere), and all these affect the transport of heat on the Earth's surface in a greater or lesser extent. These components all interact on a wide range of spatial and time scales. Atmosphere plays a key role in regulating the climate of our planet. The atmosphere does not act as an isolated system. The exchange of energy takes place between the atmosphere and other parts of the climate system, including the most important parts – the oceans. For example, ocean currents move heat from equatorial latitudes towards cooler polar latitudes. Heat is also transmitted through moisture. Water evaporating from the surface of the oceans has recorded the heat which is released later in the process of condensation, which forms clouds and precipitation. Oceans accumulate greater quantities of heat, unlike the atmosphere. Above 200 metres, the world's oceans supply 30 for about 30 times more heat than the atmosphere. Therefore, the flow of energy between ocean and atmosphere can have dramatic effects on climate change.

- Theoretical considerations about climate changes. The global climate must be considered as an option within the complex atmospheric/ocean/ice/land system (Oliver 2005). Any change in this system, resulting in climate change, creates the forcing agents causing climate change. Such forcing agents can be either internal or external. Forcing mechanisms include external forces acting outside the climate system. In contrast, internal mechanisms operate within the climate system.
- Non-radiative forcing. Any change in climate must include some forms of distribution of energy within the global climate system. However, forcing agents that do not directly affect the energy budget of the atmosphere (balance between the incoming solar radiation and the Earth's outgoing radiation) are considered as non-radiative mechanisms of global climate change. Such forces usually operate over large time scales (10^7–10^9 years) and generally include those factors that affect the climate of the geometry of the Earth's surface. Such factors are location and size of the mountain and state of the ocean basins.
- Radiation forcing. The process that changes the energy balance of the Earth-atmosphere system is known as the radiation forcing mechanism. These mechanisms include changes in the orbit of the Earth around the Sun, solar radiation, volcanic activity, and atmospheric composition. Before exploring some of the important mechanisms, internal and external, there is one factor that is also important for climate and refers to the time scale in which climate change is occurring.
- Time scale of climate change. Research on climate change is of great importance to the different time scale in which it occurs. Climate change occurs in all time

scales in terms of random and periodic forcing factors. In every period of several years to hundreds of millions of years, there is a primary reference level of random changes in climate that are caused by internal processes that are associated with feedback mechanisms, often known as Stochastic or random mechanisms. Such a coincidence involved in climate change, and its existence is due to the complex and chaotic behaviour of the climate system in response to forcing. Substantial effect on the existence of random processes is that much of climate change cannot be predicted.

- Periodic forcing factors. Periodic forcing factors of climate are far more relevant for understanding the mechanisms and influences on climate change. Unlike the random factors, periodic factors can predict future climate change. However, it is often not quite clear in what way the climate system reacts to the periodic forcing factors. If we assume that the climate system reacts to the linear way of periodic forcing, climate change will manifest similar periodicity. For the forcing of the climate, there are many factors that extend in a huge range of periodicity. The longest, 200–500 million years, is related to the passage through the solar galaxy system and changes in galaxy dust. These mechanisms are considered external forcing mechanisms. There are various external forcing mechanisms acting on time scales from 10 to 10^9 years, including galactic, solar, and orbital changes. Changes in the galaxy orbit of the solar system around the centre of galaxy are considered as a possible mechanism of the external forcing. In the mid-nineteenth century, scientist Kroll proposed a theory concerning the orbital changes that are associated with periodic changes in the Earth's orbit around the Sun. His ideas were later refined and elaborated by Milankovic (Fig. 24.1).

24.3 Milankovic Theory of Climate Change

Milankovic theory is known as the astronomical theory of climate change or Earth‘s long-term climate change. Milinkovic has identified three types of orbital changes that could act ass mechanism for forcing of climate (see Fig. 24.2):

1. **Earth eccentricity** (deviation) of the Earth's path around the Sun (100.000–400.000 years)
2. **Earth obliquity** (inclination)-range of the tilt of Earth's axis of rotation or change of the slope of the Earth's axis of rotation (~41.000)
3. **Earth precession**-revolution axis of rotation of the Earth (~21.000 years)

Although the variability of the Sun, in advance, is considered as another external factor of forcing, it remains a controversial mechanism of climate change on all time scales. Despite the many attempts to demonstrate statistical link between solar periodicity and global climate cycles, there are not any real challenging proposed mechanism to link these two phenomena. Internal forcing mechanisms are operating over time scales of 1–10^8 years. They can be either radiative or non-radiative

Fig. 24.1 Milinkovic Milutin (1879–1958)

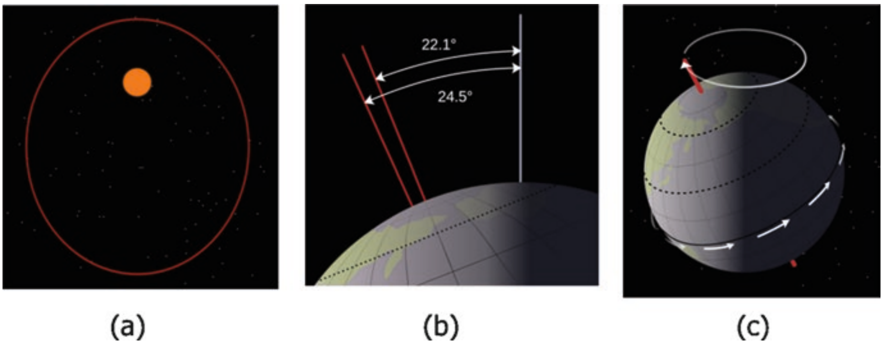

Fig. 24.2 Three cycles of Milankovic astronomical theory. (**a**) Eccentricity; (**b**) obliquity; (**c**) precession. Courtesy NASA/JPL-Caltech

forcing mechanisms. Other long-term changes 10^6 10^8 years include non-radiative forcing mechanisms.

They include continental electricity orogeny and vertical movements in the Earth's crust, affecting the sea level. These are internal mechanisms of the forcing. External changes in the amount of solar radiation and the Earth's orbit around the Sun, volcanic activity, ocean circulation, and atmospheric composition occur over all time scales from 1 to 10^5 years. Additionally, there are many other internal feedback mechanisms that contribute to changing the global climate change. The reaction of the climate system to this combination of factors forcing depends on the different times of reaction, the various components in the system. Atmosphere, snow, and ice at the surface, as well as vegetation, typically respond to a climate forcing for a period of hours to days.

The surfaces of the oceans have a reaction to climate change measured in years, while the deep oceans and mountain glaciers are changing within hundreds of years. Great Ice sheets grow and shrink over a thousand years, while parts of the geosphere respond only in forcing periods lasting hundreds and thousands of million years. The reaction of the climate system to forcing episodes could be considered as a form of resonance. When the period of forcing corresponds closely to the time of the response component of the special system, climate response will be greatest within that component.

- Climate feedback. The state of the global climate is related to general stability, which occurs as a result of the balance that exists between the common components of the global climate system. The amount of incoming solar radiation is balanced with the amount of outgoing terrestrial longwave radiation, such as the Earth continues to heat or not to cool. It is said that the Earth's climate is in equilibrium. When the climate system reacts to radiative forcing, this balance is temporarily upset, and difference between incoming and outgoing radiation appears. To establish equilibrium, the global climate changes by heating or cooling, depending on the direction of the initial forcing.
- Although the climate system is in balance, it really is dynamic and constantly changing. The system constantly adjusts to disturbances which are caused by forcing. This setup contributes to climate change. While transferring from one to another component of the system, it modifies. In some cases, the system will be intensified (positive feedback); in other cases, it can be reduced (negative reaction).
- Climate sensitivity. The concept of turning the climate response is associated with climate sensitivity and climate stability. It is useful to have a measure of the strength of various feedback processes that determine the ultimate response of the climate system to any change in radiative forcing. In general, the initial change in temperature due to the change in radiative forcing is modified through a complex combination of feedback processes.
- Empirical study of climate. To understand the present climate and predict future climate change, it is necessary to have theoretical and empirical observation. Every study of climate change includes construction (or rehabilitation) of the time series of climate data. As these data concern climate change through time, they provide a measure (quantitative or qualitative) of climate change. Climate statistics include a wide range of different types of data: temperature, precipitation, wind, humidity, evaporation, pressure, and solar radiation (light).

24.4 Climate Classification

It is assumed that the first attempt to change the classification was made by the ancient Greeks, who divided each hemisphere into three zones: torrid, temperate, and frigid. Since the beginning of the twentieth century, there were many dealings with the classification of climate, but the well-known and the most common scheme for presenting a model of change is the scheme proposed by Vladimir Köppen (1846–1940) (Fig. 24.3).

Köppen-Geiger climate classification (1980–2016) recognizes five major groups of climate, each designated by a capital letter: A (wet tropical), B (dry), C (mid-latitude wet, mild winters, moderate), D (mid-latitude wet, cold winters), and E (polar). Four groups (A, C, D, and E) are defined by air temperature. The fifth (B) group takes precipitation as the primary factor.

24.5 Climate Factors (Controllers)

Köppen's climate classification defines the global climate based on meteorological models. These models depend on factors that control and manage the atmospheric processes. The basic climatic factors are:

1. Geographic latitude (changes in the influx of solar energy and differences in temperature, largely dependent on geographical latitude)
2. Influence land/water (maritime climate is mild in principle, while continental climates are more extreme)
3. Geographic location and persistent winds (moderate effect of water is more pronounced on the upwind side of the continent)

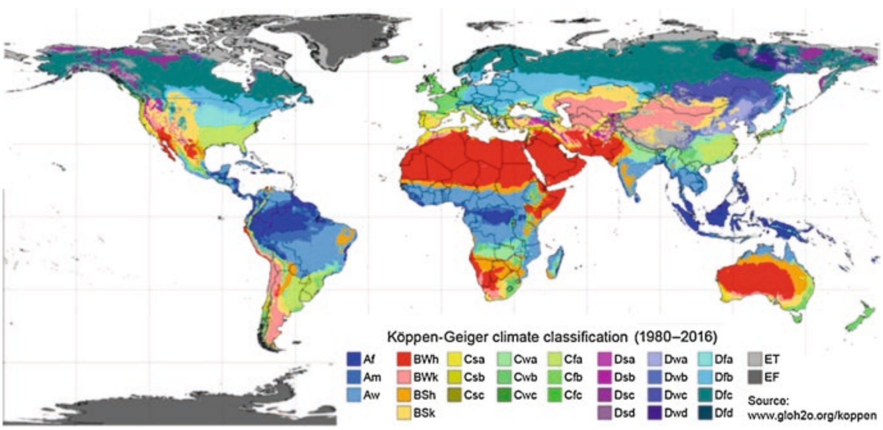

Fig. 24.3 World Köppen-Geiger climate classification (1980–2016)

4. Orography (mountain barriers prevent maritime air masses from the depth of penetration into the mainland, causing orographic rains, which create specific climate regions)
5. Ocean currents (flows moving towards the poles causing air temperatures to be higher than expected)
6. The distribution of pressure and wind (global distribution of precipitation is closely related to the distribution of the main pool of bilge systems and general wind patterns in the Earth)

24.6 Climate Types

A – Climate type is a humid tropical climate, which stretches along the equator, with constant high temperatures. In combination with precipitation throughout the year in the climate field, lush vegetation is created, known as tropical rain forest. The average temperatures in these regions, each month, usually amount to 25 °C or more, and the daily temperature changes significantly exceed the seasonal differences. Precipitation in this type of climate is normally 1750 to 2500 mm annually and is more variable than temperature, such as seasonal, as well as space. Thermal convection combined with convergence along Intertropical Convergence Zone (ITCZ) leads to a widespread rise in warm, moist, unstable air, which creates ideal conditions for the occurrence of rainfall.

B – Climate type. Dry regions in the world are covering about 30% of land area on Earth. The characteristic feature of the climate type B, i.e. the dry climate, the annual deficit is the amount of rainfall. Climatologists define climate as dry climate when annual rainfall is less than the potential loss of water through evaporation. To define the boundary between dry and wet climate, Köppen's classification applies a formula that represents three variables:

1. Average amount of rainfall
2. Average temperature
3. Seasonal distribution of rainfall

C – Climate type. Climate type C refers to the occurrence of wet climates in high latitudes. Winters are mild, so that the mean temperature of the coldest month is in the interval (−3 °C and 18 °C). There are several subgroups of C-type climate.

D – Climate type. Type D climate refers to the continental climate with wet winters expressed. The mean temperature of the coldest month varies in the interval (−3 °C to 3 °C), and the mean temperature of the hottest month exceeds 10 °C. The main annual temperature amplitudes on Earth occur in this climate type.

E – Climate type. Climate type E refers to the polar climates in which the mean temperature of the hottest month is less than 10 ° C. The annual temperature amplitudes are extreme, with the lowest annual averages of the Earth. Although polar climates are classified as wet, rainfall is usually insufficient. In some terrestrial stations, the annual amount of rainfall is even less than 250 mm. There are two types of polar climate: the tundra and the climate of perpetual ice. The first type is known as the tundra climate, which is a less forest region with grasses, sedges, mosses, and lichens with permanently frozen subsoil, called permafrost. The ice cap climate (EF) does not have a single monthly mean temperature above 0 °C. Consequently, the growth of vegetation is deprived, and the landscape is a one of permanent ice and snow.

Mountain climate. Mountain climates are characterized by large variability in climate conditions over a small area. Although the well-known climatic effects of increased height to lower temperatures, widespread rainfall is greater as a result of orographic rising. Variety and variability best describe the mountain climate. As atmospheric conditions change with altitude and exposure to sunlight, almost unlimited variety of local climates occurs in mountain regions.

24.6.1 Modern Term of Climate

The climate today is generally defined as a comprehensive statistical description of complex weather, including extreme processes over sufficiently long period, from 30 to 100 years (National Research Council 2002). There is no sharp distinction between weather and climate. There are free modes of atmospheric circulation that have a time scale to about 2 years (quasi-2-year oscillation). There are also joint ocean-atmosphere modes that have a time frame of weeks to several decades. A dominant feature is the phenomenon “El Niño-Southern Oscillation”, which occurs in the eastern tropical Pacific Ocean, with time scale of about 4 years.

24.7 Climate Modelling

Climate models attempt to simulate the behaviour of the climate system. The goal is to comprehend the key physical, chemical, and biological processes that govern climate. By understanding the climate system, it is possible to get a clearer picture of past climates, by comparison with empirical observations, and thus to predict future climate change. The models can be used for simulating the climate of different spatial and temporal scales (Warner 2011). Sometimes it is necessary to study regional climate. Other cases of interest include other weather climate models in global scale, which simulate the climate of the planet. The basic physical processes that must be treated with the preparation of climate model are:

1. Radiation processes (covering the transmission of radiation through the climate system, absorption, reflection, scattering, etc.)
2. Dynamic processes (horizontal and vertical transfer of energy, such as advection, convection, turbulence, diffusion)

3. Surface processes (including processes related to the effects of albedo on land/ocean/ice, emissivity and exchanges of energy between the Earth's surface and the atmosphere)

These processes are important for describing the behaviour of the climate system. The basic laws and other relations necessary for modelling the climate system are expressed through a series of equations that can be empirical or primitive. Solving equations is usually done by methods of finite differences. Therefore, it is important to examine the spatial and temporal resolution of the model.

24.7.1 Climate System Simplification

All models used to forecast climate, greatly simplify the complex climate system. This is due to the limited knowledge of the climate system and partly as a result of computer limitations. Simplification can be achieved in terms of spatial dimensionality, space, and time resolution or parameterization processes that simulate. The simplest models of the spatial dimension are of zero order. The state of the system is defined by a single global average. Other models always include increased dimensional complexity of 1D, 2D and 3D models. Regardless of the spatial dimension of the model, further simplification takes place in terms of spatial resolution. The time resolution of climate models substantially changed from minutes to years, depending on the nature of the models and the problem under investigation. To protect the computer stability, there should be a link between spatial and temporal resolution of the model. Sometimes atmospheric processes on scales smaller than the resolution of the model equations that describe the processes cannot be solved explicitly. In this case, their parameterization is carried out, which primarily covers the use of simplified (sometimes semi-empirical) functions or parameters. Atmospheric processes in subscale, such as turbulent processes in thunderstorms, should be parameterized because it is not possible to solve them explicitly. Other processes may also be parameterized to reduce the required computer computational time. Certain processes may be omitted from the model if their contribution is negligible compared to the time scale which is of interest. For example, there is no need to consider the role of deep oceanic circulation for modelled changes over time scales from years to decades. Some models can process radiative transfers largely but cannot ignore or parameterize energy transport. Other models can provide 3D representation, but they contain much less detailed information about the radiative transfer.

24.7.2 Climate Models

The most common division of the climate models is the following:

- Models of energy balance (EBMs)
- (1D) radiation-convective models (RCMs)

- (2D) statistical-dynamical models (SDMs)
- (3D) general circulation models (GCMs)

Patterns of energy balance are the easiest (0D) models that simulate energy balance in the atmosphere. The remaining 1D, 2D, and 3D models have gradual increase in complexity depending on the degree to which they simulate individual processes and their temporal and spatial resolution. The simplest models allow little interaction between the primary processes, radiation, dynamics, and surface processes where the most complex models are fully interactive and coupled (Trenberth 2010). Patterns of energy balance (EBMs) simulate two basic processes that govern the state of climate:

(a) Global radiative balance (between incoming solar radiation and Earth terrestrial outgoing radiation)
(b) Latitudinal transfer of energy (from the equator towards the poles)

Radiative-convective models. Radiation-convective models (RCMs) are 1D or 2D, with height as a dimension that is currently irreversible. RCMs simulate in detail the transfer of energy through the thickness of the atmosphere, including:

(a) Radiative transformations that occur when energy is absorbed, emitted, and scattered
(b) The role of convection, transfer of energy by atmospheric vertical motion, in maintaining stability

Two-dimensional Regional Climate Models (RCMs) can also simulate horizontal-averaged transfers of energy. Statistical-Dynamic Models (SDMs) are essentially two-dimensional (2D) in form, usually with one horizontal and one vertical dimension, although two versions of horizontal dimensions have been developed. SDMs combine the horizontal transfer of energy modelled by patterns of Energy Balance Models (EBMs) with radiation-convection approach of RCMs. Energy transfer from the equator to the pole is simulated in a sophisticated way, based on theoretical and empirical relations of the cell flow between latitudes. Parameters such as direction and wind speed are modelled by statistical relations, and to obtain a measure of energy diffusion, like EBMs, applying the fundamental laws of motion. General Circulation Models (GCMs) represent the most sophisticated attempt to simulate the climate system. Three-dimensional (3D) formulation of the model is based on fundamental laws of physics:

(a) Conservation of mass
(b) Conservation of momentum
(c) Conservation of energy

Series of primitive equations describing these laws are solved, summarizing assessment of the wind field, which is expressed as a function of temperature.

Processes such as cloud formation also simulate. To calculate the basic atmospheric variables at each network point, the use of large computer time is required, because of the large number of points and therefore expensive computer GCMs. 3D models can provide a reasonable representation of the planetary climate and, unlike the simple models, may simulate processes in global and continental scales (such as the effects of mountain ranges of the atmospheric circulation). However, many of the GCMs are unable to simulate synoptic (regional) weather phenomena such as tropical storms, which play an important role in latitudinal transfer of energy and the amount of movement. Spatial resolution of GCMs is also limited in vertical dimension. Consequently, many processes in the border layer must be parameterized. The new generation models are transitional-coupled atmosphere-ocean GCMs, as an attempt to accurately simulate the climate system (Bretherton et al. 1990; Gates et al. 1992). Such coupled models created major difficulties for computer components of the atmosphere and ocean over the corresponding different time scale. As a summary, GCMs cannot reasonably be regarded as simulating global climate and climate in continental scale but lacks reliability in relation to regional scale.

24.8 Climate Change

24.8.1 Scientific Overview of Climate Change

Science has made enormous inroads in understanding climate change and its causes and is beginning to help develop a strong understanding of current and potential impacts that will affect people today and in coming decades (Burroughs 2007). This understanding is crucial because it allows decision-makers to place climate change in the context of other large challenges facing the nation and the world. There are still some uncertainties, and there always will be in understanding a complex system like Earth's climate. Nevertheless, there is a strong, credible body of evidence, based on multiple lines of research, documenting that climate is changing and that these changes are in large part caused by human activities. While much remains to be learned, the core phenomenon, scientific questions, and hypotheses have been examined thoroughly and have stood firm in the face of serious scientific debate and careful evaluation of alternative explanations. As a result of the growing recognition that climate change is under way and poses serious risks for both human societies and natural systems, climate change represents one of the most challenge scientific and interdisciplinary topics and the global concern as it broadly affects a wide range of human and natural systems posing significant and vulnerable risk impacts.

24.8.2 Climate Change Factors

From climatology it is known that in the past, climate of the Earth changed in wide limits. There were times when hot climate prevailed much of the planet without an ice layer on the poles, climate, and ice, when most of the moderate zone was under a thick layer of ice. Climate change caused three types of factors:

(a) Astronomy
(b) Planetary
(c) Anthropogenic

Influence of astronomical factors is manifested in a very long time, tens of thousands of years. Therefore, they cannot be considered when considering climate change over a century. Planetary phenomena which can lead to a change of climate on Earth are the big volcanic eruptions and the fall of large meteors. When these phenomena occur, large quantities of aerosols are disposed into the stratosphere. These particles remain very long in the high layers of the atmosphere and cause solar radiation to weaken and thus contribute to less heat travelling to the Earth's surface. Another important effect of volcanic eruptions causes a discharge of large quantities of carbon dioxide in the atmosphere. In the last century, when a global increase in the temperature of the Earth's atmosphere was noticed, planetary phenomena that could have caused such warming were not registered. Therefore, these factors should be excluded from the current deliberations on climate change. We continue with the consideration of the anthropogenic factors. They include industrialization, the cutting of forests, urbanization, irrigation, and processing of land. The more frequent use of fossil fuels, the spread of atomic power plants, and the increase of atmospheric haze of the various aerosols have already received global scale. Climate changes reflect variations that occur in the Earth's atmosphere and the processes in other parts of the Earth, such as oceans, ice belts, and the effects arising from human activities. However, all listed factors have local information on climate change, with the exception of industrialization, whose stormy development in the last century caused the excretion of large amounts of carbon dioxide (CO_2) and other greenhouse gases (methane, nitrogen oxide, chlorofluorocarbons, water vapour), to create a "greenhouse effect". Besides the effect of "greenhouse", soon other anthropogenic influences can be expected from climate change on the Earth. Global warming, as a result of human influence, is more expressed at the poles, where additional danger arises from the gradual melting of large icy expanses. On the other hand, across the dry area and the area around, the Intertropical Convergence Zone (ITCZ) in turn moves to the north, so that there are dangers of expanding and deepening droughts to northern latitudes. These climate change and climate doubts are expected to reach global scale, with serious impact on people's lives, material goods, and the economy of the entire civilization and sustainable development. Rapid globalization, technological development, industrialization, and uncontrolled emission of greenhouse gases are the main factors for the occurrence of a change of energy balance in global scale, a change that contributes to the distortion of the general atmospheric circulation and the appearance of increased frequency of extreme in the atmosphere (such as droughts, floods, tornadoes, local eddies with great energy, strong shots of wind, snow drifts).

24.9 Climate Variability

The atmospheric processes have a wave nature with a different spatial scale, from millimetres, as a small turbulent eddy to kilometres such as the global planetary waves, and time scale from seconds to years. While *weather* deals with the current atmospheric status or atmospheric conditions quantitatively represented through distribution of the main meteorological parameters (e.g. temperature, moisture, pressure, rainfall) in particular area and time, *climate* describes an averaged pattern of weather over longer period of time in respect to temperature and rainfall distribution, usually over 30 years (normal climate period) or very long-term evidence. In order to clarify the complex term of climate change, we must understand climate variability. According to Heinlein we usually say, "Climate is what we expect, and weather is what we get". One way to understand the difference between weather, climate variability, and climate change is to think about how they operate on different time scales. As it is shown on Fig. 24.4, weather and climate have a different time scales from hours, days (weather), months, and years to decades (climate variability), through centuries (climate change). While the term climate change implies long-term climate change, climate variability represents climate (variations) deviations or inconsistencies (Lau and Waliser 2012).

There are different patterns of variability: recurrent temperature patterns or other climate variables. They are quantified with different indexes. The indexes are

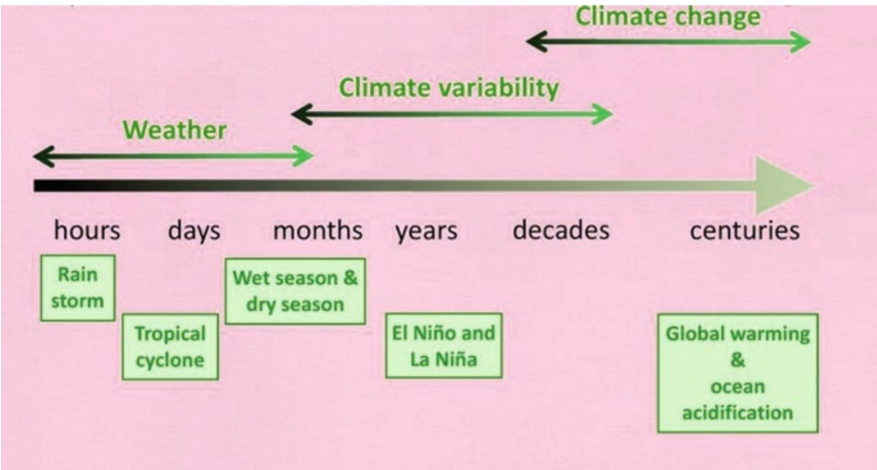

Fig. 24.4 Weather and climate scales

characterized by simplicity and completeness, and each index usually represents the status and time of the climate factor it represents. By their nature, the indices are simple and combine many details into a generalized, complete description of the atmosphere or oceans that can be used to characterize the factors influencing the global climate system. Climate oscillations or climate cycles are any repetitions within a global or regional climate. These fluctuations in air temperature, sea surface temperature, precipitation, or other parameters can be quasi-periodic, which often occur at interannual, multi-year, centennial, millennial, or longer periods. A typical example is El Niño (South Oscillation, which includes surface temperatures of the equatorial Pacific and western tropical South America) which affects the climate of the entire planet. The artic amplification and the role of the Solar activity on natural climate variability will be discussed in the following chapters.

24.9.1 Arctic Amplification

Climate scientists increasingly point out to the impact of the potential connections between Arctic warming and extreme weather across the mid-latitudes. The Arctic is warming more quickly than the global surface average (see Fig. 24.5). This phenomenon is known as "Arctic amplification". In part, this stems from the rapid loss of sea ice cover in the region. As Arctic sea ice diminishes, energy from the

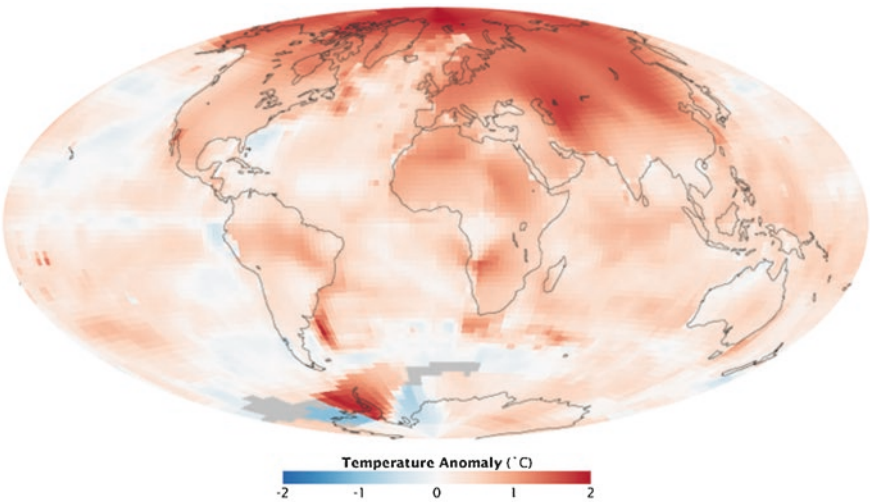

Fig. 24.5 Global temperature anomaly. (Credit: NASA image by Robert Simmon, based on GISS surface temperature analysis data including ship and buoy data from the Hadley Centre. Caption by Adam Voiland)

Sun that would have been reflected away by the bright white ice is instead absorbed by the ocean, causing further warming. (Declining snow cover over Arctic land areas has the same effect.) Arctic amplification is also caused by temperature feedbacks. As the Earth's surface warms, it emits more energy back to space. But less energy is radiated back from the Arctic compared with lower latitudes, meaning the region warms more quickly. While a warmer Arctic Ocean further inhibits sea ice growth, it also generates warmer and moister air masses over the Arctic and nearby continents. A warming Arctic also reduces the temperature difference with the mid-latitudes, which has consequences for circulation patterns in the atmosphere (more on this later).

The second theory refers to "sudden stratospheric warming", when the stratosphere around the Arctic region is in the process of strong warming. The temperature of the middle of the stratosphere, above the pole, is more than 65 °C warmer than in the normal strongly polar vortex conditions (low-pressure spinning area). This helps to weaken the polar vortex and may even cause it to fall into two or more smaller vortices. This sudden stratospheric warming as a process on a larger scale had implications in terms of weather and strengthening and persistence of winter conditions in the middle/late January, in much of Europe.

24.9.2 Solar Activity

At the end of this analysis, before giving a summary of the above, let's not forget that according to estimates by scientists dealing with solar radiation, for the next 11-year solar cycle, predicted reduced activity of the Sun, like the current one (Roy 2018). The current 24-h solar cycle shown in Fig. 24.6 is decreasing, and according to NASA's calculations, it is projected to reach a sunny minimum – a period when the Sun is at least active – at the end of 2019 or 2020. "The effects of cooling that we see in the Earth's thermosphere are the result of the current solar minimum conditions". The thermosphere is a layer of the Earth's atmosphere, starting about 100 km above the Earth's surface and highly sensitive to solar activity. "There is no connection between the natural cycle of cooling and warming in the thermosphere and the influence of weather / climate on the surface of the Earth".

NASA and other climate researchers continue to record a trend of warming the troposphere, a layer of the atmosphere closest to the surface of the Earth. There is inconsistency between the scientific indicators for heating the troposphere (global warming) and the thermosphere. Some scientists claim that the cooling of the thermosphere beyond 100 km above the Earth's surface will have an impact on the troposphere and will cause a "mini ice age". These claims are, however, denied based on reports of record low temperatures as part of a natural cycle and reduced sunshine activity, which is not in line with the current scientific knowledge of the global warming of the troposphere. The identification of these oscillations becomes possible after significant improvement of the global observations and developing extensive series of daily pressure and geopotential height data. Different modes of these

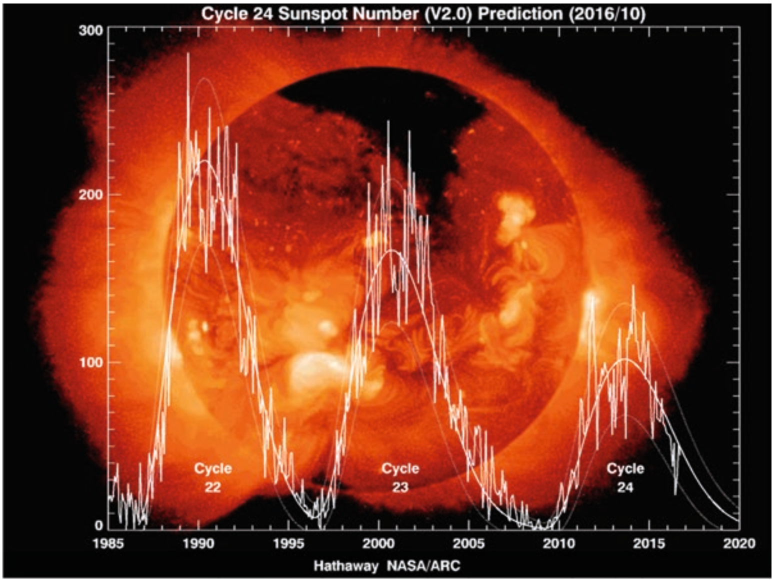

Fig. 24.6 The solar cycle. Courtesy NASA/JPL-Caltech

oscillations can persist for months and years having a major influence inter-seasonal or annual regional and global climate, modifying storm tracks and the distribution of climatic anomalies. Ocean currents play a major role in maintaining the heat balance on the Earth. Besides creating surface currents, winds may also cause vertical movements of water, such as cascades of cold water, and can replace warmer surface water from deeper layers.

24.10 Climate Change Ingredients

There are many ingredients that affect climate on Earth, such as the amount of greenhouse gases and aerosols in the atmosphere and the characteristics of the Earth's surface. These factors determine how the solar energy input is retained or reflected into space. The concentration of greenhouse gases in the atmosphere of CO_2, CH_4, and N_2O is significantly increased since the industrial revolution. This is largely due to human activities such as the combustion of fossil fuels, cutting forests, agricultural activities, changes in the natural environment, and others. There are numerous observations that indicate the successive increase in air temperature and the temperature of the oceans, the global melting of snow and ice, and the raising of the sea level. According to statistics in the last 100 years, global air temperature has increased for about 0.74 °C, and global sea levels have increased to 17 cm, partly due to melting of snow and ice from the mountains and polar regions. Besides these, other changes have been observed, including changes in Arctic temperatures, changes in salinity of the ice and oceans, and a change in the movement of winds, droughts, rainfall, warm waves of frequency, and intensity of tropical cyclones. In the next two decades, the temperature is expected to grow 0.2 °C per decade. The continuous emission of gases will cause further increases in global temperatures, hence, other climate changes are expected in the 21st century. If the warming continues over the following centuries, it will lead to ice cap melting, thus raising the global sea level of approximately 7 metres. In the twenty-first century, a greater impact from climate change is expected, which will affect natural systems. If there is a significant warming, the ability of ecosystems to adapt will disappear, and the risk of extinction of species will increase.

24.10.1 Greenhouse Effect

Greenhouse effect is the heating of the Earth due to the presence of greenhouse gases. Shortwave radiation from the Sun passes through the Earth's atmosphere, and it is absorbed by the Earth's surface, causing it to heat. Part of the absorbed radiation energy is then reemitted back to the atmosphere, and it is known as longwave infrared radiation. One part of the infrared waves can pass through the greenhouse gases in space, while another part remains trapped by them. Greenhouse gases absorb the waves and emit them back towards Earth, causing further warming of the Earth's surface. It is believed that global warming is the result of increased greenhouse effect, which in turn is due to the increased concentration of greenhouse gases in the atmosphere. The main greenhouse gas created by human is carbon dioxide (CO_2). Larger quantities of CO_2 in the atmosphere increase the average level of outgoing radiation. Besides carbon dioxide, man produces methane and nitrogen oxide, which has a lesser amount. CO_2 covers 75% of the total emission of greenhouse gases in the world. Emission of greenhouse gases that covers all releases of this type of gases in the atmosphere is the result of smoke, steam, and smog from exhaust gases, chimneys, fire, and other sources. Most of the carbon dioxide is released by burning fossil fuels, which are a widely used energy source.

24.10.2 Global Warming

Global warming is an increase in the global average air temperature, at the Earth's surface and oceans, from the mid-twentieth century until today (Philander 2008; Mathez and Smerdon 2018). Experiments with global models of the atmosphere have shown that if the current concentration of carbon dioxide in the atmosphere has increased by 40%, the mean annual temperature in the lower latitudes would dramatically increase by about two degrees. The consequences of it would be moving the subtropical belt towards higher latitudes, with more frequent and longer droughts, sudden melting of sea ice cover in the polar regions, and the increase in intensity and frequency of weather disasters in many areas of our planet. These predictions about the general climate changes that will contribute to global warming are quite authentic, and some of them have already been registered in recent years. However, more precise estimates cannot be given yet, because of two reasons: first, the lack of multiplication of mathematical models of the atmosphere, whose validity decreases with the length of the integration, and the second reason is that we still do not know well enough the so-called feedback processes that are characteristic for the climate. Warming from anthropogenic emissions from the pre-industrial period to the present will persist for centuries to millennia and will continue to cause further long-term changes in the climate system. The synthesis report on climate change as contribution from the working groups to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC 2014) delineates a new scientific evidence of climate change, observational analyses, theoretical studies of climate processes, and climate model simulations and future climate projections reveals that the changes and the warming of the climate systems are unambiguous. The report summarizes that global warming of the Earth's atmosphere in the second half of the last century had increased about 1.0 °C above pre-industrial levels, with a likely range of 0.8 °C to 1.2 °C and likely to reach 1.5 °C between 2030 and 2052 if it continues to increase at the current rate with high confidence. Climate-related risks for natural and human systems depend on the magnitude and rate of warming, geographic location, levels of development and vulnerability, and on the choices and implementation of adaptation and mitigation options (high confidence). This summary addresses the following significant topics:

- Observed changes and their causes
- Future climate change, risks, and impacts
- Adaptation, mitigation, and sustainable development

24.11 Current Climate Change Status

Human influence on the climate system is evident, and recent anthropogenic emissions of greenhouse gases are the highest in history with widespread impacts on human and natural systems.

(a) Since the 1950s warming of the climate system is definite. The atmosphere and the ocean have warmed, the amount of snow and ice have reduced, and sea level has risen, and anthropogenic greenhouse gas emissions have increased since the pre-industrial era.
(b) Changes in climate have caused impact on natural and human systems. Impacts are due to observed climate change, irrespective of its cause indicating the sensitivity of natural and human systems to changing climate.
(c) Changes in many extreme weather and climate events have been evidenced since 1950. Some of these changes are associated with human influences, including a decrease in cold temperature extremes, an increase in warm temperature extremes, an increase in high sea levels, and an increase of several heavy precipitation events in many regions.

24.12 Future Climate Projections, Risks, and Impacts

Continued emissions of greenhouse gasses will cause further warming and long-lasting changes in the all components of the climate system.

(a) Cumulative emissions of CO_2 largely determine global mean surface warming.
(b) Surface temperature increase over the twenty-first century under all assessed scenarios.
(c) The amounts of snow and ice have diminished.
(d) Snow and ice cap melting, global sea level rising, warming, and acidification of ocean.
(e) Increased frequency of weather and climate events (e.g. heat waves and the extreme precipitation events in many regions).

24.12.1 Adaptation, Mitigation, and Sustainable Development

Adaptation and mitigation represent the supplementary measures to tackling climate change in order to prevent the negative impacts of climate change and reducing vulnerability and risk impacts to the effects of climate change:

(a) Effective decision-making to limit climate change and its effects by a wide range of analytical approaches for evaluation of the expected risks.
(b) Without additional mitigation efforts and adaptation, warming by the end of twenty-first century will lead to very high risk of severe, widespread, and irreversible global impacts.
(c) Taking a longer-term perspective in the context of sustainable development will enhanced the future options and preparedness for reduction of the risks of climate change impacts.
(d) Multiple mitigation pathways to limit warming to below 2 °C.

These pathways require substantial emission reduction over the next few decades and *near zero emissions of* CO_2. Many adaptation and mitigation options can help addressing climate change, but no single approach is enough by itself. Effective implementation depends on interdisciplinary approach, policies, cooperation, and measures at all scales. Mitigation options are available in every major sector, in a more cost-effective way, using an integrated approach that combines measures to reduce energy use and the greenhouse gas emissions. Climate change is a threat to sustainable development. There are many opportunities to link mitigation, adaptation, and need of other societal objectives through integrated responses. Climate change is a reality, but its impact on precipitation, evaporation, surface runoff, and flood risk is still uncertain. The other effects of global warming include melting of ice (Fig. 24.7), extinction of species, and increase of the spectrum of diseases and the withdrawal of glaciers.

Glaciers are large masses of ice moving over land, formed by dense snow in the area where the snow is accumulating than melting and sublimating (Fig. 24.8). Because glaciers are sensitive to changes in climate, they are indicators of global warming. Availability of water plays a significant role in these influences of air temperature. In the context of contemporary research on climate change, the European Space Agency (ESA) carried out two missions in duration of 3 years. The first mission is to measure the water content in the soil around the planet every 3 days to a depth of 2 m.

This will be followed by photosynthesis and growth of plants, which are crucial to the calculation of the process in which carbon dioxide is released and absorbed, especially by plants and oceans. The second mission is to measure changes in the content of salt on the surface of the seas, which will improve understanding of the basic engines of global patterns of oceanic circulation. Changes in salinity of ocean water depend on the surplus or deficit of fresh water through evaporation and precipitation and, in the polar regions, and the melting of freezing ice. To conduct this research, on November 2nd, a satellite (SMOS) was launched from the Kosmodrom in Plesetsk, northern Russia. Its aim is to research the impact of climate change on water movement through land, air, and seas. Being the first cosmic instrument for measuring the Earth's water and soil salinity of the oceans, scientists hope SMOS will be able to fill important gaps in our knowledge about the importance of planetary vital driven cycle.

Fig. 24.7 Effects of climate change. (Credit: Jason Auch, https://creativecommons.org/licenses/by/2.0/deed.en)

Fig. 24.8 Patagonia glacier. (Credit: Luca Galuzzi. I, Luca Galuzzi [CC BY-SA 2.5 (https://creativecommons.org/licenses/by-sa/2.5)])

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Chapter 25
Meteorological Measurements and Observations

Why are atmospheric observations necessary? Geophysical and associated meteorological observations, and monitoring of the natural environment, are carried out for various reasons. They are used for the preparation of weather analysis in real time, the forecasts and warnings of natural disasters related to weather, water, and climate. Meteorological measurements and observations are used for the local operational activities that depend on time, such as flight-route operations at airports, construction works, transportation, agricultural activities, health, and others. Meteorological and climate data are important for research on climate and climate change. Meteorology has made significant progress in the quality and diversity of weather forecast services since the launch of the first meteorological satellites in 1957/1958 gave rise to the World Weather Watch (WWW) in 1963. Growing global temperature will cause a rise in sea level and is expected to increase the intensity of extreme weather events and to change the amount and mode of precipitation. But current societal challenges – due to the unfolding impacts of climate change – demand further evolution of the Earth observation network: an upgrading of the global space- and surface-based observing systems (Fig. 25.1) and the adoption of a new and integrated approach that incorporates recent scientific and technical advances.

25.1 Meteorological Observing Systems

The requirements for observational data may be met using:

- In situ measurements
- Remote sensing (including space-borne) systems

The Global Observing System, designed to meet these requirements, is composed of the surface-based subsystem and the space-based subsystem. The surface-based subsystem comprises a wide variety of types of stations according to the

V. Spiridonov, M. Ćurić, *Fundamentals of Meteorology*,
https://doi.org/10.1007/978-3-030-52655-9_25

Fig. 25.1 Liquid-in-glass thermometers

application (e.g. surface synoptic station, upper air station, climatological station, etc.). The space-based subsystem comprises a number of spacecraft's with on-board sounding missions and the associated ground segment for command, control, and data reception.

The representativeness of an observation is the degree to which it accurately describes the value of the variable needed for a specific purpose. For instance, synoptic observations should typically be representative of an area up to 100 km around the station, but for small-scale or local applications, the considered area may have dimensions of 10 km or less. According to Orlanski (1975), horizontal meteorological scales may be classified as follows, with a factor two uncertainty:

(a) **Microscale** (less than 100 m) for agricultural meteorology, for example, evaporation
(b) **Local scale** (100–3 km), for example, air pollution, tornadoes
(c) **Mesoscale** (3–100 km), for example, thunderstorms, sea and mountain breezes
(d) **Large scale** (100–3000 km), for example, fronts, various cyclones, cloud clusters
(e) **Planetary scale** (larger than 3000 km), for example, long upper tropospheric waves

25.2 Basic Meteorological Elements

The present state and the further behaviour of the atmosphere are defined using the basic meteorological elements such as the temperature, pressure, and humidity of the air, wind speed and direction, cloud cover, precipitation, and visibility (the transparency of the atmosphere), as well as soil and surface water temperatures, solar radiation, and longwave terrestrial and atmospheric radiation. They also include weather phenomena such as thunderstorms and snowstorms. The variations in the meteorological elements are the result of atmospheric processes, and they determine the weather and the climate. The meteorological elements are observed at aerological and meteorological (weather) stations and at meteorological observatories by means of aerological and meteorological instruments. The following elements are observed at a station making surface observations:

- Present weather
- Past weather
- Wind direction and speed
- Cloud amount
- Cloud type
- Cloud-base height
- Visibility
- Temperature
- Relative humidity
- Atmospheric pressure
- Precipitation
- Snow cover
- Sunshine and/or solar radiation
- Soil temperature
- Evaporation

When we know these items, others such as dew point, freezing point, and other elements can be defined as "secondary elements". There are instruments for measuring these elements, except for the type of clouds.

25.3 Standardization of Measurements

Many of the elements necessary for synoptic, climatologic, or aeronautical targets can be measured automatically. Observations and measurements of meteorological elements are made in the meteorological and weather stations on certain standardized procedures and methods in order to be consistent, representative, and uniformed (e.g. Brock and Richardson 2001; Emeis, 2010; Harrison, 2015). Each state, through its National Weather Service, establishes and maintains its own national network of stations. Data from these measurements, according to the standards and procedures of the World Meteorological Organization (WMO), are sent on international exchange as a part of the global telecommunications surveillance system. The term "standard" and other similar expressions denote different instruments, methods, and steps used to avoid uncertainties in the measurements. To effectively control the standardization of meteorological instruments of national and international scale, the WMO has adopted a system of national and regional standards. WMO has adopted a system of national and regional standards. Most of the elements required for synoptic, climatological, or aeronautical purposes can be measured by automatic instrumentation. All the instruments should also be carefully calibrated and regularly maintained. Trained observers are required and/or certified by an authorized meteorological service to establish their *competence to make observations* to the required standards. They should have the ability to interpret instructions for the use of instrumental and manual techniques that apply to their own observing systems. Meteorological observing stations are designed so that representative measurements (or observations) can be taken according to the type of station involved.

Siting and exposure. The following considerations apply to the selection of site and instrument exposure requirements for a typical synoptic or climatological station in a regional or national network:

(a) Outdoor instruments should be installed on a level piece of ground, preferably no smaller than 25 m x 25 m where there are many installations.
(b) The ground should be covered with short grass or a surface representative of the locality and surrounded by open fencing or palings to exclude unauthorized persons.
(c) Within the enclosure, a bare patch of ground of about 2 m x 2 m is reserved for observations of the state of the ground and of soil temperature at depths of equal to or less than 20 cm.
(d) There should be no steeply sloping ground in the vicinity, and the site should not be in a hollow.
(e) The site should be well away from trees, buildings, walls, or other obstructions.
(f) Very open sites which are satisfactory for most instruments are unsuitable for rain gauges.
(g) If in the instrument enclosure surroundings, maybe at some distance, objects like trees or buildings obstruct the horizon significantly, alternative viewpoints should be selected for observations of sunshine or radiation.
(h) The position used for observing cloud and visibility should be as open as possible and command the widest possible view of the sky and the surrounding country.
(i) At coastal stations, it is desirable that the station commands a view of the open sea. However, the station should not be too near the edge of a cliff because the wind eddies created by the cliff will affect the wind and precipitation measurements.
(j) Night observations of cloud and visibility are best made from a site unaffected by extraneous lighting.

Symbols, units, and constants. Instrumental measurements produce numerical values. The purpose of these measurements is to obtain physical or meteorological variables representing the state of the local atmosphere. For meteorological practices, instrumental readings are variables, such as "atmospheric pressure", "air temperature", or "wind speed". For meteorological observations, the following units are used:

(a) Atmospheric pressure (p), expressed in hectopascals (hPa)
(b) Temperature (T), expressed in degrees Celsius (°C) or (T) in Kelvin (K)
(c) Wind speed, at the surface and upper level observations, in metre per second (m $s^{-1)}$
(d) Wind direction expressed in degrees clockwise from the north or in the scale (0–36), where 36 is the wind from the north and 09 the east wind (°)
(e) Relative humidity (U), expressed in percent (%)
(f) Precipitation (total amount) expressed in millimetres (mm) or (kg $m^{-2)}$

25.4 Measurement of Air Temperature

The traditional instruments used for measurement of air temperature are called thermometers. Meteorological thermometers commonly in use can be widely classified into following categories based on the physical processes made use of in their design.

Basic standards in measurement. In meteorological practices, the surface air temperature refers to the temperature near Earth's surface at a height of between 1.25 m and 2 m above the ground level. In order for the air temperature to be representative and reliable, it is necessary for the area around the meteorological station to be in free external conditions and a large enough (spacious) area, and it should not be disturbed by some external influences or by the surrounding objects. For agro-meteorological practices, the surface air temperature measurements are generally required to be made at different level near the ground.

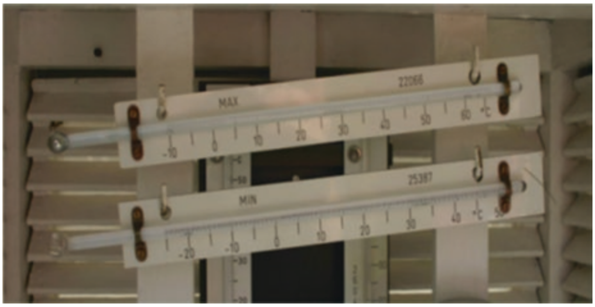
Liquid-in-glass thermometers. These are thermometers (see Fig. 25.1) that are based on the principle of the expansion of liquid in a fine bore glass tune to measure temperature. The most broadly used among these are commonly known as "ordinary" thermometers which are employed to measure prevailing atmospheric temperature. Others suitably modified in their design, which are used to indicate extreme temperatures attained during the period they are exposed to the atmosphere, are called as "maximum" or "minimum" thermometers (Fig. 25.2).

Two types of liquid-in-glass thermometers commonly in use are:

(i) Mercury-in-glass thermometers
(ii) Spirit-in-glass thermometers

Liquid-in-glass thermometers. Such thermometers make use of the expansion/contraction of liquid in a scaled metal container to measure air temperature. The variations in pressure in the metal container caused due to the changes in the liquid volume gives a measure of air temperature to which the container is exposed.

Fig. 25.2 Maximum and minimum thermometers



Mercury in steel and Bourdon tube thermometers fall under this category. These thermometers are essentially a pressure gauge calibrated in terms of temperatures.

Ordinary (station) thermometers. Ordinary (station) thermometers are the most precise instruments of all meteorological thermometers. Usually, such an instrument is a type of glass thermometer filled with mercury.

Maximum thermometers. A maximum thermometer mainly used for measuring air temperature is a mercury-in-glass thermometer having a constriction near the bulb end. It has a small area where the glass tube is narrower on one end. This is known as a "constriction". As the air temperature rises, the mercury in the thermometer expands and moves freely up the tube (past the constriction) until the maximum temperature occurs. When the air temperature begins to drop, the constriction prevents the mercury from flowing back down the tube. The mercury will not move back down the tube until the thermometer is shaken.

Minimum thermometers. When it comes to minimum thermometers, the most acceptable instrument is the alcohol thermometer. It contains an index of dark glass, submerged in alcohol, with a length of about 2 cm. As part of the air remains in the tube of alcohol thermometer, it should provide a protective chamber in the upper end, which should be long enough to allow the instrument to withstand a temperature of about −50 °C without being destroyed.

Ground thermometers. For measuring the temperature of the ground, at depths of 20 cm or less, mercury thermometers with glass are used. The body of this kind of thermometer is curved, with dust, or other suitable angle, below the lowest scale. The tube of the thermometer is placed in the ground at the desired depth, but the scale is read directly from the thermometer (Fig. 25.3).

Mechanical thermograph. Mechanical thermograph which is used daily works on sensors or bimetal Burdon's tube because they are relatively cheap, reliable, and flexible. However, mechanical thermometers are not suitable for remote or electronic recording of air temperature.

Bimetallic thermometers. These thermometers use a bimetallic strip/coil as a temperature-sensing element. The change in the curvature of the strip/coil, with changes in temperature, gives a measure of air temperature. At the bimetal thermograph, movement of registered pen (marker) is controlled by the change in the curve of bimetal spring or bar, which is firmly fixed on the shoulder which is set on the

Fig. 25.3 Ground thermometers

frame. To be able to change the zero of the instruments, it is a tool for fine-tuning of the shoulder.

Thermograph with Burdon's tube. The basic composition of the thermographs with a Burdon's tube is like bimetal thermograph, except that its temperature-sensitive element is in the form of a curved metal tube with a flat, elliptical section filled with alcohol. Burdon tube is less sensitive than bimetal element and usually requires a mechanism of multiplication to obtain enough scale value. The typical time constant is about 6 m, the air speed of (5 cm s^{-1}).

Electrical thermometers. This type of thermometer (Fig. 25.4) makes use of the electrical characteristics of the sensing device that generates an output signal which varies with changes in temperature. Electrical instruments for measuring air temperature are widely used in meteorology. Their principle of operation is essentially based on their ability to provide an external signal that is suitable for use in remote indication, labelling, preservation, or transmission of temperature data. Commonly used sensors are electrical resistance elements, semi-conductive thermometers (thermostats) and thermopares.

The two main types of thermometers under this category are the platinum resistance thermometer and thermistor thermometer.

Electrical resistance thermometers. This type of thermometers shown in Fig. 25.5 uses the principle of measuring the electrical resistance of the material. The manner in which the electrical resistance of a particular material changes is known in physics. Here, the representation of the temperature can be used to change the resistance. Here, the representation of the temperature can be used to change the resistance. For small temperature changes, the increase in the resistance of pure metal is proportional to the change of temperature, as given by (Eq. 25.1):

$$R_T = R_0\left[1+\alpha\left(T-T_0\right)\right] \tag{25.1}$$

Fig. 25.4 Electrical thermometer. Credit: Harke/CC BY-SA (https://creativecommons.org/licenses/by-sa/3.0)

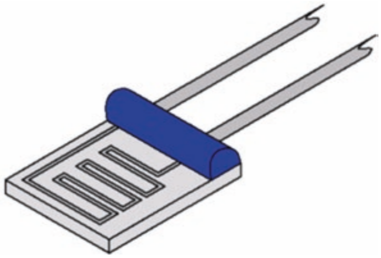
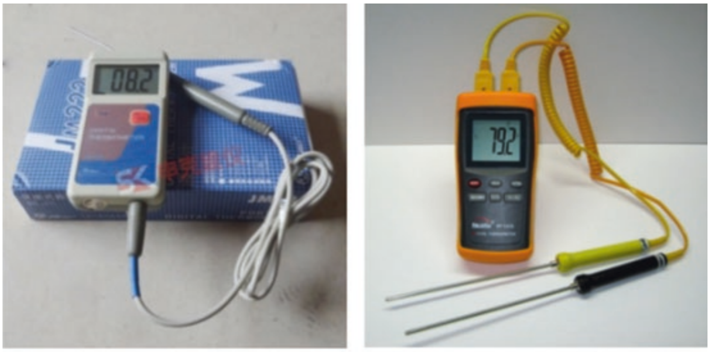


Fig. 25.5 Electrical resistance temperature

where $(T - T_0)$ is the temperature difference, R_T – resistance of a fixed amount of metal at temperature T, R_0 –resistance at reference temperature T_0 and is the temperature coefficient of resistance near $T_{0.}$

*Semi-conductive thermometers. These are a*nother type of resistive element that is widely used in thermostats. As it is shown in Fig. 25.6a, it has a semiconductor with a relatively large temperature coefficient of resistance, which can be either positive or negative depending on this material. Stalactites mixtures of metal oxides are suitable for making practical thermostats which usually have the form of small discs or spheres and are often glassy. Typical thermostats have a resistance which changes by a factor of 100 or 200 in the temperature range - 40 to 40 °C.

Thermocouples. These thermometers (Fig. 25.6b) use the principle of establishing the contact of electromotor force is used, on the place where two different metals touch. If you make a simple circle with two metals and by merging the same temperature, then there will be resultant electromotor force in the circle, because of the contact between the two metals, and the electromotor force will be reversed and the connection interrupts. If the temperature of one junction is changing, both electromotor forces will be more in balance, then we will have a total electromotor force



(a) (b)

Fig. 25.6 (**a**) Semi-conductive thermometer; (**b**) thermocouple

set up in the circle, and then the electricity will flow. The physical principle used for the construction of thermometers is known as thermocouples.

25.5 Measurement of Atmospheric Pressure

For meteorological purposes, atmospheric pressure is generally measured with electronic barometers, mercury barometers, aneroid barometers, or hypsometers. The latter class of instruments, which depends on the relationship between the boiling point of a liquid and the atmospheric pressure, has so far seen only limited application and will not be discussed in depth in this publication. In standard use for measuring, atmospheric pressure instruments called barometers are common. There are several standard types of barometers: mercury, electrical, and metal.

Mercury barometers. There is a growing trend of gradual disposal of the use of mercury barometers (Fig. 25.7), because it is highly toxic; free mercury is corrosive to the aluminium alloys used in air frames (for these reasons there are specific regulations in some countries that proscribe handling or carriage of mercury barometers). The barometer is very delicate and difficult to transport.

The instrument must be read and corrections to be applied manually. Also available are other types of sensors for pressure, with equivalent accuracy and stability with the electronic reader.

Electronic barometers. Most of the recently designed barometers use electronic assemblies that transform into electric sensor response magnitude associated with the pressure in the form of analogy or digital signals. Analogy signals can be shown in different electronic metres. Monitors and systems for collecting data, such as those used in automatic weather stations, are often used to display digital outputs or analogy to digital outputs. The barometer with current digital technology uses

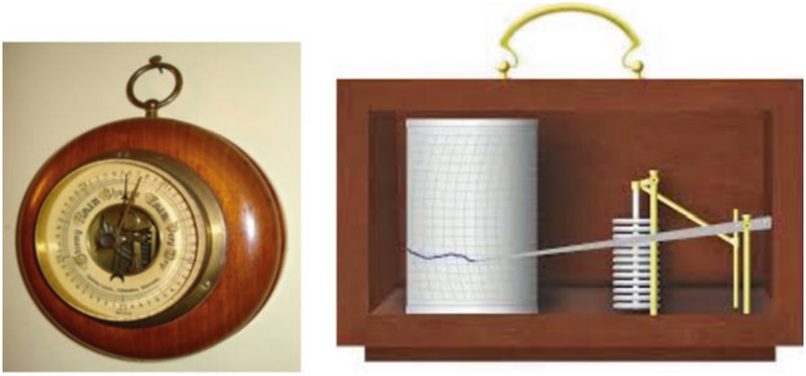
Fig. 25.7 A mercury barometer

different modules in order to improve long-term stability and accuracy of measurements.

Bourdon tube barometers. Bourdon tube barometers usually consist of a sensor element that, as for an aneroid capsule, changes its shape under the influence of pressure changes (pressure transducers) and a transducer that transforms the changes into a form directly usable by the observer. The display may be remote from the sensor. Precise and stable digital instruments with quartz Bourdon tubes are used as working reference barometers in calibration laboratories.

Aneroid barometers. The advantages of conventional barometers regarding mercury barometers are their compactness and portability, which makes them especially suitable for use at sea and in the field. The main components of the metal barometer shown in (Fig. 25.8a) are closed metal chamber, completely or partially vacant, and strong elastic system which protects the chamber from destruction under the influence of external atmospheric pressure. At any given pressure, there is a balance between the force caused by spring and that by external pressure. Steel chamber can be made of materials (steel or beryllium copper) that have elastic properties as a result of which the chamber can act as a spring itself.

Tools are needed to detect and display changes in the deviation that occurs. This can be a system of levers which are intensified deviations and moving the cursor over the scale graduated to indicate pressure. Alternatively, the light beam can be removed over the scale. Instead of these mechanical analogue techniques, certain barometers are equipped with a manual operating micrometre counter which indicates the pressure directly in the tens of hectopascals. The reading takes place when



(a) (b)

Fig. 25.8 (**a**) Aneroid barometer; (**b**) barograph

the luminescent indicator signals that micrometre only contacted metal. This type of metal is movable and robust.

Barographs. Barographs represent an aneroid barometer used for continuous recording of the changes in atmospheric pressure over time (Fig. 25.8b). It is a recording aneroid barometer. The reading of the barometric pressure over time is produced on a paper or a foil chart, which is known as barograms. The working of the barograph is very scientific, and it indicates the atmospheric pressure. The output is presented through a continuous graph on paper or foils. The barograph has a metal cylinder, which is linked with a pen arm. The pen arm is directly proportional to the changing atmospheric pressure that enables the meteorologists to study the forthcoming climate. Pressure versus weather is traced by the pen and is recorded on the chart, which is rotated by clockwork. The barograph is used both on the ships as well as on the land. Today they come in various shapes and sizes from small ones to very large ones. To give it a fancy look, the barographs even come in glass boxes. The drawer below collects the barograms. The barograph generally provides a reading for a week. For synoptic purposes they have the following characteristics:

(a) Graduation in (hPa)
(b) Readability up to (0.1 hPa)
(c) Measurement factor (10 hPa) in (1.5 cm) of the bar

25.6 Measurement of Air Humidity

The measurement of atmospheric humidity, and often its continuous recording, is an important requirement in most areas of meteorological activity. This chapter deals with the measurement of humidity at or near the Earth's surface. There are different methods for measuring the humidity of the air. Humidity of air can be expressed in several ways:

- Relative humidity
- Absolute humidity
- Water mixing ratio
- Water vapour

All these definitions of humidity can be set with the same instruments; only the procedure for obtaining a value from another course is different. Relative humidity is a relation between the current amount of water vapour in the air and the maximum amount which that same air may receive, not reaching saturation. Relative humidity is expressed in (%).

Hygrometer. Any instrument for measuring humidity is known as a hygrometer. The physical principles most widely employed for hygrometry are:

(i) The gravimetric method
(ii) Condensation method
(iii) The psychrometric method
(iv) Sorption method

The gravimetric method. This method uses the absorption of water vapour by a desiccant from a known volume of air (gravimetric hygrometer. The gravimetric method yields an absolute measure of the water vapour content of an air sample in terms of its humidity mixing ratio. This is obtained by first removing the water vapour from the sample. The mass of the water vapour is determined by weighing the drying agent before and after absorbing the vapour. The mass of the dry sample is determined either by weighing or by measuring its volume.

Condensation method. The first type is chilled mirror method (dew point or frost-point hygrometer). The basic principle is when moist air at temperature T, pressure p, and mixing ratio r_w (or r_i) is cooled, it eventually reaches its saturation point with respect to water (or to ice at lower temperatures) and a deposit of dew (or frost) can be detected on a solid non-hygroscopic surface.

Other type is heated salt solution method (vapour equilibrium hygrometer, known as the dew cell). The equilibrium vapour pressure at the surface of a saturated salt solution is less than that for a similar surface of pure water at the same temperature. The temperature of the solution at which the ambient vapour pressure is reached provides a measure of the ambient vapour pressure. For this purpose, a thermometer is placed in good thermal contact with the solution.

The psychrometric method. The psychrometer consists essentially of two thermometers exposed side by side, with the surface of the sensing element of one being covered by a thin film of water or ice and termed the wet or ice bulb, as appropriate. The sensing element of the second thermometer is simply exposed to the air and is termed the dry bulb. This is the most widely used method and is described in detail.

Sorption methods. Certain materials interact with water vapour and undergo a change in a chemical or physical property that is sufficiently reversible for use as a sensor of ambient humidity.

The psychrometer. In order to determine air temperature and humidity, a psychomotor is used (Fig. 25.10a). It consists of two thermometers, one of which (the dry

(a) (b)

Fig. 25.9 (**a**) The psychrometer; (**b**) the Assmann aspirated psychrometer

bulb) is an ordinary glass thermometer, while the other (wet bulb) has its bulb covered with a jacket of clean muslin which is saturated with distilled water prior to an observation. The less moisture in the air, the stronger the evaporation, so the difference between damp and dry thermometer is greater. From these two readings of the thermometers from the table or by calculating the formulas, all the above forms of humidity and dew point are given.

The Assmann aspirated psychrometer. The Assmann aspirated psychrometer is composed of two glass mercury thermometers mounted vertically (Fig. 25.9b). The aspirator may be driven by a spring or an electric motor. One thermometer bulb has a well-fitted muslin wick which, before use, is moistened with distilled water. Each thermometer is located inside a pair of coaxial metal tubes, highly polished inside and out, which screen the bulbs from external thermal radiation. The tubes are all thermally insulated from each other.

Screen psychrometer. Two mercury-in-glass thermometers are mounted vertically in a thermometer screen. The diameter of the sensing bulbs should be about 10 mm. One of the bulbs is fitted with a wet-bulb sleeve, which should fit closely to the bulb and extend at least 20 mm up the stem beyond it. If a wick and water reservoir are used to keep the wet-bulb sleeve in a moist condition, the reservoir should preferably be placed to the side of the thermometer and with the mouth at the same level as, or slightly lower than, the top of the thermometer bulb. The wick should be kept as straight as possible, and its length should be such that water reaches the bulb with sensibly the wet-bulb temperature and in enough (but not excessive) quantity.

The hair hygrometers. Any absorbing material tends to equilibrium with its environment in terms of both temperature and humidity. The water vapour pressure at the surface of the material is determined by the temperature and the amount of water bound by the material.

Any difference between this pressure and the water vapour pressure of the surrounding air will be equalized by the exchange of water molecules. The hair hygrometers (Fig. 25.10) utilize the absorption properties of hair. These instruments are used to measure relative humidity, and they operate on the principle of

Fig. 25.10 Hygrometer

absorption of moisture from organic substances (e.g. human or horsehair, etc.) that are changing the humidity of the air. Hygrograph is an instrument that records the relative humidity at the time and records it on paper or it memorizes it.

Heated psychrometer. The principle of the heated psychrometer is that the water vapour content of an air mass does not change if it is heated. This property may be exploited to the advantage of the psychrometer by avoiding the need to maintain an ice bulb under freezing conditions. Heat psychrometers use the principle of sustainability of the water vapour content of the mass air unit when it is heated. This feature can be used as an advantage in making psychrometer in order to avoid holding the ice tank in freezing conditions. The air is removed in the tube where it passes through an electric heat element and then comes the measurement chamber, which consists of dry and wet thermometer from the tank with water. The heat element controls circulation and provides air temperature not to be declined below a certain level, which is typically 10 °C. There are also hygrometers which are using absorption of electromagnetic radiation. Water molecules absorb electromagnetic radiation in the range of wavelengths. This feature can be used to obtain a measure of molecular concentration of water vapour or gas. The most useful belts of the electromagnetic spectrum for this purpose are the ultraviolet and the infrared belt. The method applies measurements of attenuation of radiation in the wave zone, which is typical for absorption of water vapour, along the path between the radiation source and the receiving device.

25.7 Measurement of Surface Wind

Direction and speed of wind are very important meteorological elements. Surface wind is usually measured by a wind vane and cup or propeller anemometer. When the wind blows constantly from one direction, it is called dominant or prevailing wind. Wind speed is often measured by an instrument known as anemometer. Wind is a vector quantity, which is fully determined when knowing both its components: direction and speed. The direction of the wind is part of the horizon where the wind blows, and the speed is the elapsed time of air particle pass in unit time. Direction shall be indicated by the parties of the world or the azimuth in degrees from 0 to 360, while the speed is measured in metres per second (m/s) or kilometres per hour (km/h). Because of the lack of devices for measuring speed, it is estimated by Bofor scale. The wind direction is determined by wind vane (Fig. 25.11). It is an easy indicator in the form of a shaft, mounted on a vertical axis that rotates freely around its axis. The direction of the arrow to the wind, the back has a vertical plate which serves as a "rudder". The speed of wind is measured with an anemometer (wind gauge). It is a vertical axis with 3 or 4 hollow semi-spherical (Robinson Cross) that rotates under the influence of wind.

The wind is stronger, faster spheres spin, and rotation is easily transformed into mechanical or electrical equivalent graduate units for wind speed. Instead of semi-spherical, sometimes a small propeller is used. Anemometer is used for measurement of the mean velocity in a period (usually 2 or 10 minutes) or current speed. Measurement is performed on standard height of 10 metres above the surface

Fig. 25.11 Wind vane

(ground). Anemograph measures the speed of wind at the time and the measured data recorded on paper or stored electronically.

25.8 Measurement of Precipitation

Precipitation is the most important meteorological element that quantifies the amount of fallen rainfall at the ground. Precipitation is defined as liquid or solid products of condensation of water vapour, which falls on the Earth's surface from the clouds. Precipitation includes several forms of hydrometeors: rain, hail, snow, dew, frost, freezing rain, and tiny droplets of mist. The total amount of rainfall that falls at the ground in a period of time is expressed in the form of vertical thickness, water depth or snow water equivalent, in case of solid forms. Snow cover is also expressed through the thickness of the fresh, fallen snow covering a horizontal surface. Unit rainfall is linear thickness, usually expressed in millimetres (volume/area) or (kg/m$^{2)}$ (mass/area) of liquid precipitation. That means, for example, 1 mm = 1litre/cm^2. The daily measurements of precipitation should be taken at fixed times common to all network or networks of interest. Rainfall of less than 0.1 mm is generally known as a trace. The amount of rainfall (intensity) is similarly expressed in linear measure per unit time, usually in millimetres per hour. Measurements of snow precipitation are taken in units and tens of centimetres to the nearest 0.2 cm. Lower value of 0.2 cm is generally called a trace. The thickness of snow on the Earth's surface is usually measured daily and expressed in centimetres.

Instruments for measurement of precipitation. Generally, an open receptacle with vertical sides is used, usually in the form of a right cylinder, with a funnel if its main purpose is to measure rain. The instrument for measurement of precipitation is called rain gauge (Fig. 25.12a). This instrument is a container in the shape of the

Fig. 25.12 (**a**) Rain gauge, (**b**) totalizer, (**c**) pluviograph

cylinder, placed vertically with a hole on top and with an area of 200 cm^2. Rain comes through the hole, which is mixed in a vessel at the bottom. Around the vessel there is a closed space for thermal insulation to prevent evaporation of accumulated precipitation. During the reading, accumulated water in the vessel is discharged in a tube with scale in millimetres and quotes its quantity. Rain gauge usually sets the column, the height of one (1) metre above the Earth's surface. For measurement of precipitation in inaccessible areas where observations are not continuous, the most common instruments are totalizers (Fig. 25.12b). Besides rainfall, its intensity can also be measured. The intensity of rain is expressed in millimetres per minute (mm/min). The intensity of rainfall is measured with an instrument called *pluviography* (Fig. 25.12c).

Measurement of precipitation of snow and snow cover. Snowy precipitation is the thickness of the freshly fallen snow, deposited at the Earth's surface in a specified period (usually 24 h). In this way, snowfall does not include deposition of moved or blown snow. For purposes of measurement of thickness, the word "snow" should also include ice pallets, graupel or hail, and plate ice, which appear to be formed by precipitation directly or indirectly.

Average depth of snow is the total thickness of snow on the Earth's surface, at the time of observation. Water equivalent of snow cover is the vertical thickness of the water that would be obtained by melting of the snow cover. Direct measurements of the thickness of snow cover in the open country are made with engraved stick or scale (Fig. 25.13). In order for representative thickness measurements to be obtained, numerous vertical measurements in places covered by snow need to be

Fig. 25.13 Measurement of snow thickness

made. The stick should be placed on a flat surface that is not leeward, nor covered with some of the surrounding objects (tree, building facilities, etc.). The total amount of snow cover is measured, but only the snow that fell in the last 24 hours.

25.9 Measurement of Solar Radiation

The Sun and other elements emit a spectrum of electromagnetic radiation, which depends on its temperature. The temperature is higher, so the spectrum is strongly shifted to the side of higher frequencies. But for the meteorological purposes, two variables are measured from solar radiation: the duration of radiation on any point on the Earth's surface in a certain period (day, month, year) and energy that comes from the Sun, a certain area at some time.

Heliograph is an *instrument* for recording the duration and intensity of sunshine (Fig. 25.14). This type of recorder is made up of a glass ball which concentrates sunshine on to a thick piece of card. The sunshine then burns a mark on the card which shows the number of hours of sunshine in the day.

Pyranograph. The energy that a surface receives from the Sun is usually measured by pyranograph or pyradiograph (Fig. 25.15a) and is expressed in Julies per metre square per hour (J/cm^2 h$^{-1)}$, or some other proportional units. It should be noted that these instruments must be in an open space, which is not covered with object such as trees, buildings and the like. In these objects, of course, mountain barriers are not included.

Pyrheliometer. The direct solar radiation is measured using pyrheliometer (Fig. 25.15b). This means the instrument is always aimed directly at the Sun, via a tracking mechanism that continuously follows the Sun. It is sensitive to wavelengths in the band from 280 to 3000 nm. This device is composed of metal surfaces that receive radiation and are placed perpendicular to the solar radiation.

Angstrom's compensation pyrheliometer works on the principle of two manganese plates: one is heated by the Sun, Moon, and other electrical paths, until they

Fig. 25.14 Heliograph

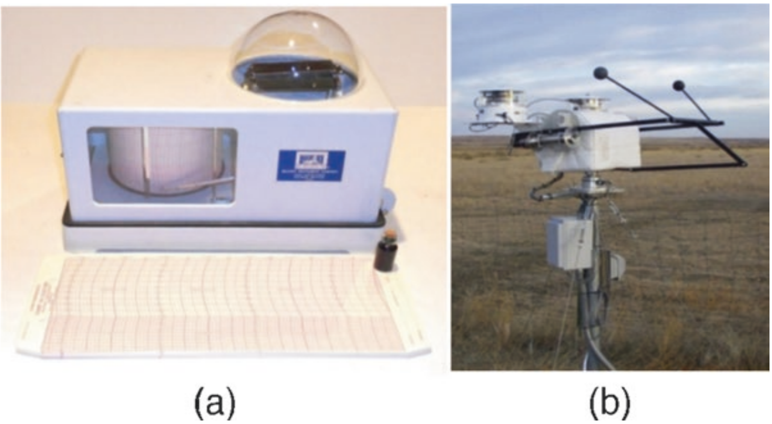


Fig. 25.15 (**a**) Pyranograph; (**b**) Pyrheliometer

Fig. 25.16 Pyranometer

reach an identical temperature. Electricity required for heating the plate is proportional to incoming shortwave radiation.

Pyranometer. Pyranometer is an instrument used for measurement of the solar radiation at full angle, the flat surface, and the diapason of the spectrum of radiation from (300 to 3000 3000 nm). This instrument is sometimes used to measure solar radiation on inclined surfaces in a horizontal position and turned to reflected measured global radiation (Fig. 25.16). When the diffusive component of solar radiation is measured, a component of direct solar radiation is protected from the pyranometer with a device for darkening. Pyranometers are normally used as thermal power sensors, photoelectrical, pyroelectrical, or bimetal elements. Since pyranometers are constantly exposed to all weather conditions, they must be solid in design and resistant to external influences and effects of moist air, especially near the sea.

Measurement of longwave radiation from the Earth can be achieved either indirectly, by subtracting the measured global radiation, measured by the total radiation, or directly, using pyrgeometres. Frequently pyrgeometres eliminate short wavelengths using filters that have constant visibility for long wavelengths, while they are almost non-transparent for short wavelengths (300 to 3000 nm). Longwave radiation, shortwave radiation, and total radiation are measured radiometers representing two instruments or a hybrid instrument ventilation polyethylene dome and a carefully balanced detector response.

25.10 Measurement of UV Radiation

Measurement of solar UV radiation is necessary, primarily because of the negative effects that this has caused to the environment and human health. These measurements are of interest for monitoring the growth of radiation on the Earth's surface, due to ozone depletion. UV spectrum is conventionally divided into three parts:

(a) (UV-A) radiation with a range of wavelengths in the interval from 315 to 400 nm), just outside the visible spectrum. It is less biologically active, and its intensity on the Earth's surface does not change the content of atmospheric ozone.
(b) (UV-B) radiation in the belt of wavelengths from 280 to 315 nm. This radiation is biologically active. Its intensity depends on the atmospheric ozone column. A frequently used expression for its biological activity is its negative skin effect, which means the extent to which radiation causes redness of the human skin.
(c) (UV-C) radiation at wavelengths from 100 to 280 nm is completely absorbed in the atmosphere and does not occur naturally on the Earth's surface.

The most sophisticated commercial instruments for measuring UV are called "radiometers" (Fig. 25.17). They use holographic traps to disperse incoming energy spectrum. Low-energy UV radiation compared to that of the visible spectrum requires a strong blocking from outside the belt. This is achieved by using double monochromatic or blocking filters, which transmit only UV radiation. The measurement of output from monochromatic commonly uses fotoamplifier.

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Fig. 25.17 Instrument for measurement radiation (radiometer)

Fig. 25.18 Transmissometer

25.11 Measurement of Visibility

Meteorological visibility is an element that is very important in the measurements. Its size can be estimated by the observer. Assessment of visibility depends on many subjective and physical factors. Meteorological size, which is transparent to the atmosphere, can be measured objectively, and it represents the meteorological optical range. Visibility can be measured with telephotometric instruments, which are referred to us as transmissometers (see Fig. 25.18) that are designed for daily measurement of the ratio of attenuation, by comparing the visible light a separate facility to the facility of the sky. But they are usually not used for routine measurements because of the preferred use of direct visual observations.

Transmissometers represent instruments that are commonly used for measuring the average coefficient of attenuation. This instrument consists of a transmitter that provides modulated flux light from a source with constant average power, and receiver, i.e. accompanied photodetector. Frequently used sources of light are halogen lamps. Modulation of the light source prevents disturbance of sunlight. The transmission factor is determined by the output of photodetector, which allows calculating the ratio of attenuation and meteorological optical distance.

25.12 Measurement of Evaporation

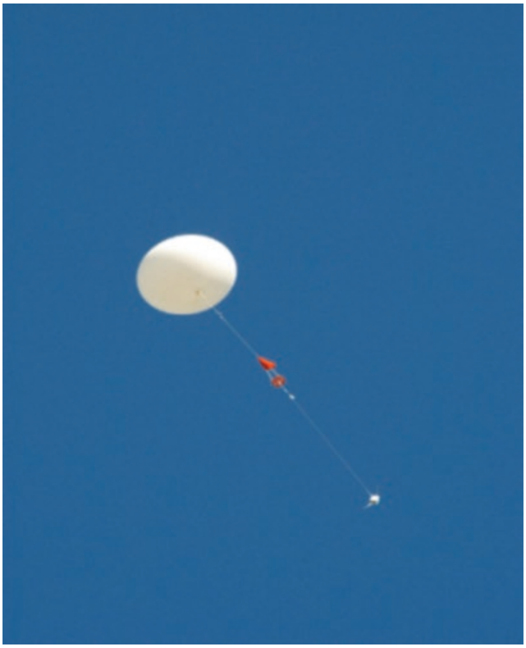
Evaporation represents the quantity of evaporated water from open water surface or from the ground. Transpiration is the process through which water from the plants is transferred to the atmosphere in the form of water vapour. Hence, evapotranspiration (or effective evapotranspiration), means an amount of water vapour that evaporated from the Earth's surface and plants, in terms of natural moisture content. Potential evaporation is the quantity of water vapour, which can be broadcasted from the surface of pure water from unit surface area and unit time, of the existing atmospheric conditions. Potential evapotranspiration is a maximum amount of water that can evaporate in each climatic region, the continuous expanse of vegetation. Speed of evaporation is defined as the amount of water evaporated from unit area in unit time. It can be expressed as mass or volume of liquid water, evaporated per unit area in unit time, usually as an equivalent depth of liquid water, evaporated in unit time from the whole area. Direct measurements of evaporation or evapotranspiration, the widespread natural water or land surfaces, are not practically feasible at the present time. However, several indirect methods are developed that are performed by point measurements or other calculations that provide reasonable result. The water lost by default saturated surface is measured with evaporimeters (Fig. 25.19), which can be classified as atmometers, and evaporimeters the container or tank. Atmometer is an instrument which measures the loss of water from the wet porous surface. Moist surfaces are porous ceramic spheres, cylinders, containers, or exposed filter paper discs saturated with water. The evaporating element of atmometer is consisting of a ceramic sphere (~5 cm) in diameter, associated with reservoir of bottle with water through glass or metal tube. Subsequent measurements of the volume of water that remains in the engraved tube will give the amount lost by evaporation at any given time.

25.13 Upper Air Measurements

For the measurement of meteorological elements, pressure, temperature, and humidity at a certain height, a device called radiosonde is used (Fig. 25.20). Radiosondes consist of a balloon into the atmosphere, equipped with devices for measuring one or

Fig. 25.19 Evaporimeter

Fig. 25.20 Radiosonde



more meteorological parameters. It is provided with a radio transmitter, which is used for sending information to the observation station. Radiosondes can attach to the balloon or may be dropped from an airplane or rocket. The stations, at which measurements are conducted, are made from electronic tools. Upper measurements of temperature and relative humidity are used to initialize the analysis of numerical forecast models for operational weather forecasting. Radiosondes provide most of the measurements of temperature and relative humidity directly above the land, while radiosondes launched from separate islands or vessels have limited coverage over the oceans. Temperatures, with vertical scale like radiosondes, can be measured by air at different levels of cruises. Aircraft measurements are used as an addition to upper air sounding measurements, especially overseas. Satellite measurements of temperature and distribution of water vapour have lower vertical resolution than radiosondes or aircraft measurements. Satellite measurements have the greatest impact on the analysis of numerical forecast over oceans and other areas of the globe where upper air sounding and aircraft measurements are rare or not available. Precise measurements of the vertical structure of the field and temperature profile of water vapour in the troposphere are extremely important for all types of forecast, especially for the regional and local forecasting. The vertical structure of the field of temperature and water vapour determines the stability of the atmosphere. Radio sounding measurements are vital for studies referred to the pollution of the environment and climate change with height.

25.13.1 Measurement of Upper Wind

Upper winds data are mainly obtained by using radio sounding techniques, although when additional data are needed for elevation winds without cost to launch the radiosonde can use pilot balloons (Fig. 25.21), which use the principle of determining upper level winds with optical recording of the free balloon and radio wind measurements (determination of upper winds with recording the free balloon by electronic tools). Measurements of upper stations in the Global Observer System over land are supplemented with measurements from aircrafts, wind profilers, and Doppler weather radar. The system for measuring the wind profile, known as LIDAR, provides a measurement of the direction and speed of wind above ground level to about 5 km altitude.

25.14 Advanced Remote Sensing Measurements

The requirements for observational data may be met using in situ measurements or remote sensing (including space-borne) systems, according to the ability of the various sensing systems to measure the elements needed. Thus, a different type of measurements and observations is continuously carried out to monitor the weather, climate, and water of our planet. Some of the measurements have a special research

Fig. 25.21 Pilot balloon

nature in frame of some research programme, field campaign, experiment, or other activities:

- Measurements and observations at aeronautical meteorological stations
- Aircraft observations
- Marine observations
- Special profiling techniques for the boundary layer and the troposphere
- Rocket measurements in the stratosphere and mesosphere
- Locating the sources of atmospherics
- Satellite observations
- Radar measurements
- Balloon techniques
- Urban observations
- Road meteorological measurements

25.14.1 The Global Observing System

The Global Observing System is designed to meet these requirements and is composed of the surface-based subsystem and the space-based subsystem. The surface-based subsystem comprises a wide variety of types of stations according to the application (e.g. surface synoptic station, upper air station, climatological station, and so on). The space-based subsystem comprises several spacecrafts with on-board sounding missions and the associated ground segment for command, control, and data reception.

25.14.2 LIght Detection And Ranging (LIDAR)

LIDAR is a remote sensing method that uses light in the form of a pulsed laser to measure ranges (variable distances) to the Earth. These light pulses combined with other data recorded by the airborne system generate precise three-dimensional information about the shape of the Earth and its surface characteristics. A LIDAR instrument (see Fig. 25.22) principally consists of a laser, a scanner, and a specialized GPS receiver. Airplanes and helicopters are the most commonly used platforms for acquiring LIDAR data over broad areas. There are two types of LIDARs: topographic and bathymetric. Topographic LIDAR typically uses a near-infrared laser to map the land, while bathymetric LIDAR uses water-penetrating green light to also measure seafloor and riverbed elevations. LIDAR systems allow scientists and mapping professionals to examine both natural and manmade environments with accuracy, precision, and flexibility. Over sea, the winds at altitude are mainly obtained by civilian aircraft at different flight levels. In future, LIDAR data from satellites and radars is expected to be used in order to obtain wind measurements and improve global coverage of current monitoring systems.

25.14.3 SOund Detecting And Ranging (SODAR)

SODAR is an instrument used for remote measurement of dispersion of sound waves caused by disturbances in the atmosphere. It is adjusted to measure the speed of wind at different heights and layers just above the ground (Fig. 25.23). These types of remote sensing systems have several advantages. SODAR operate a single, simple process and include coverage of greater height. The main advantage of the system is that it can be installed in a very short time.

Fig. 25.22 LIDAR

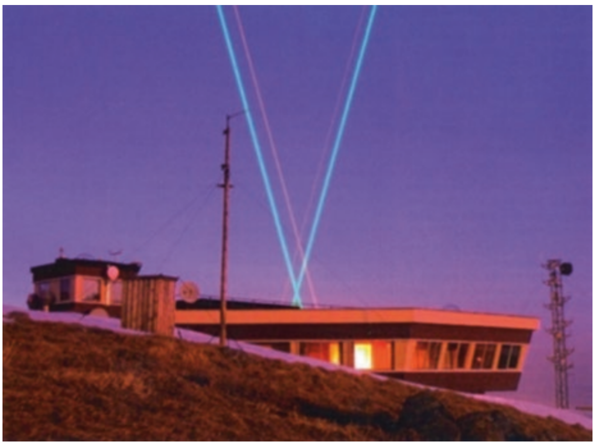


Fig. 25.23 SODAR

The standard height limit of the meteorological towers is (~150 m), where measurements with SODAR exceed this limit and provide a precise reading. Radio sounding methods for measuring the speed and direction of wind at altitude essentially depends on the monitoring of movement or free balloon, which rises to a uniformed rate or by placing the object under the influence of gravity. If you need to measure a given horizontal movement of air, there should not be a significant horizontal motion in terms of air that is monitored. The core information needed by systems directly includes the targeted amount and measurements of its flat position or, alternatively, its horizontal speed at certain intervals. Remote measurement systems are used to measure the movement of the atmosphere with the rejection of electromagnetic radiation or sound, of one or more of the following purposes: hydrometeors, sand, aerosols, etc. the index breaches caused by atmospheric turbulence at small scales or in the air molecules themselves.

25.15 Measurements at Automatic Weather Stations (AWS)

Automated weather station (AWS) is defined as a meteorological station where measurements are made of meteorological elements and are transferred automatically (Fig. 25.24). Instrumental measurements are read from the outside by means of sensors. The measured data is then disseminated to the central unit for acquisition of data. Data collected by measuring devices can be processed locally at the present

Fig. 25.24 Automated weather station – AWS

automatic station or elsewhere, for example, the central processor in the network. Automatic weather stations can be designed as an integrated concept of different devices in combination with measurement units for reception and processing of data. They are used to meet several needs, ranking from the simple assistance to the observer to complete replacement of the observations in completely automatic stations.

25.16 Satellite Observations

Application of techniques of remote control in the measurement and monitoring of meteorological variables and supporting the Earth's surface, troposphere, and stratosphere using satellite are called satellite measurements (Webber 2009; Tan, 2016). Satellite systems continue to evolve, and for several years, new systems are expected to be operational. Typical meteorological satellites orbit the Earth at heights of about (36,000 km) or approximately (850 km), and they are used for obtaining images and quantitative information about the properties of the surface and lowest (20 km) from the atmosphere (Fig. 25.25). The use of sensors with satellite platforms has its advantages and flaws in terms of measurements, compared to the use of surveillance systems from the Earth surface. Ability to obtain images from meteorological satellites is part of the justification for the use of satellites. Cloudiness images provide valuable diagnostic information in the support analysis

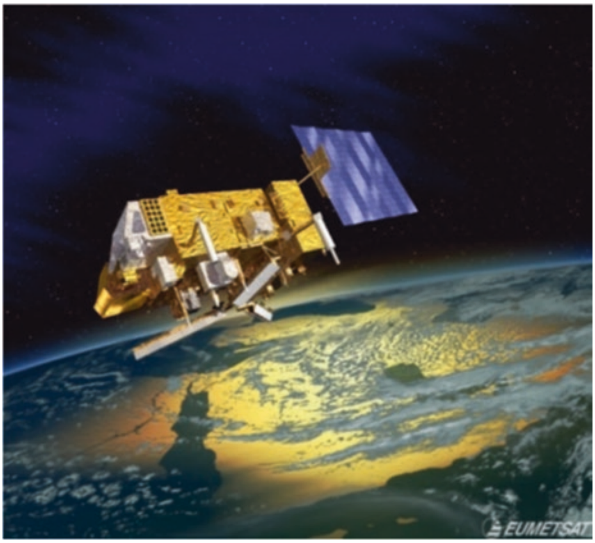


Fig. 25.25 Meteorological satellite

of meteorological characteristics. Meteorological variables are measured operationally in the present time, with variable resolution and accuracy, including:

- Profile of temperature and temperature at the top of the cloud at the surface of the sea and land
- Profile of moisture
- Wind at the cloud and the ocean surface
- Liquid and total water intensity of rainfall
- Total radiation and albedo
- Cloud type and height of cloud tops
- Total ozone
- Coverage and height of the ice and snow

With the help of satellites, non-meteorological variables such as vegetation, volcanic dust, etc., which are operationally important, are also measured.

Meteorological satellites of the second generation are designed to monitor the advantage of new technologies and improve the already successful and proven design of the original "Weather-Sat" satellites. SEVIRI radiometer (MSG-2) satellite represents a total of 12 channels that generate images through scanning the Earth every 15 minutes. Visible channel provides the highest resolution data at 1 km resolution; other channels provide data at 3 km. This generation of satellites is a powerful tool for obtaining precise information about the state of the atmosphere in real time.

Besides obtaining data on cloudy systems in different parts of the globe, many applications for measuring and monitoring atmospheric phenomena and processes with the help of satellite are developed.

Today they have successfully detected the development of cyclones and other atmospheric disturbances of smaller scale. Satellites are successfully used for prediction and monitoring of development and evolution of tropical cyclones, hurricanes, typhoons, and tornadoes. Numerous applications for prediction of fires, air pollution, the emergence of low stratus clouds or fog, solar eclipse, and related atmospheric phenomena are also developed. Meteorological Satellite Third Generation (MTG) is a six-satellite system of four imaging satellites carrying imaging and lightning detection mission as well as two sounder satellites providing infrared and ultraviolet capabilities, for both climate and meteorological applications. It will provide the atmospheric chemistry and air quality information.

Radar observations and measurements. The word radar comes from the English term "Radio Detection and Ranging". It operates on the principle of radio waves that are emitted in sequences and refuse to back the cloudy particles. The time between emission and detection can compute the distance of the object. The intensity of the signal depends on the concentration of facilities, size of particles, and their type (snow, city, or rain). Radar data is usually shown as reflexivity in decibels. The Doppler radar (Fig. 25.26), using speed mode, is based on the Doppler effect (see Doviak and Zrnic 2006), which can determine the direction and speed of the object (precipitation). The most prominent meteorological radar today is Doppler dual-polarization radar. They improve the accuracy of precipitation estimates, leading to better flash flood detection, ability to discern between heavy rain, hail, snow, and sleet, and improve detection of non-meteorological echoes (e.g. ground clutter, chaff, anomalous propagation, birds, and tornado debris) and identification of the melting layer (e.g. bright band).

Radar measurements are very important in operational meteorology and forecast the weather, especially in the very short-term forecast and the announcement of

Fig. 25.26 Radar measurements (Doppler radar)

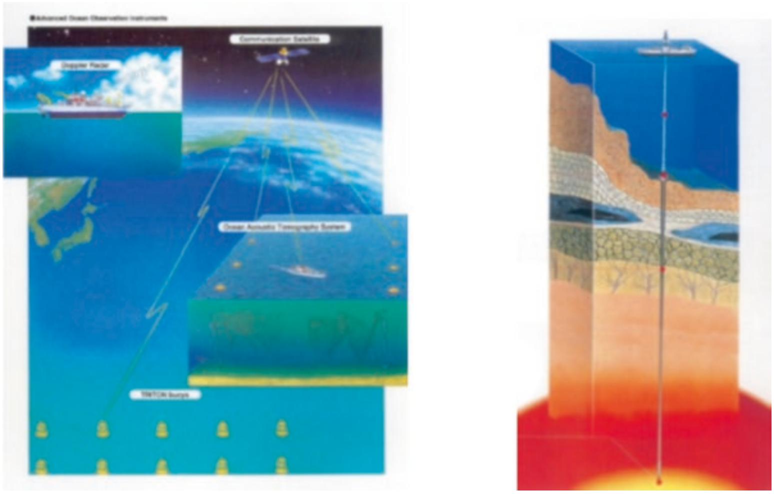


Fig. 25.27 Advanced Observation Instruments on "CHikyu"

weather disasters (nowcasting), related storms, strong winds, electrical discharges, heavy rainfall, and the hail occurrence.

25.17 Special Measurements

Ocean drilling in the twenty-first century. For meteorological, and much general, for geophysical needs, the Chikyu platform was developed with a programme of special measurements in the atmosphere, in deep oceans, and deep into the land beneath the ocean. "Chikyu" will enable to drill 7 km under the sea floor at maximum water depth more than 4 km (see Fig. 25.27). This extremely complex ship is specially built to serve as a mobile platform for special measurements that should provide answers to many unknown geophysical issues, such as why the occasional quasi periodically occurs the worming of ocean water in a tropical deck of the Pacific. This has a significant influence on the weather and climate of not only these areas but also much wider.

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