MODULE 1: METEOROLOGY

BANGLADESH METEOROLOGICAL DEPARTMENT

LOCAL TRAINING

Project:
Strengthening Meteorological Information Services and Early Warning System (Component-A)

Prepared by:
Grant Thornton Consulting Bangladesh Ltd.

AIMS AND OBJECTIVES:

- Participants can demonstrate skills for interpreting and applying atmospheric observations.
- They can demonstrate knowledge of the atmosphere and its evolution.
- Participants can demonstrate knowledge of the role of water in the atmosphere.
- Participants can demonstrate facility with computer applications to atmospheric problems.
- Participants can demonstrate skills for communicating their technical knowledge.

DELIVERY AND DESCRIPTION:

Methodology:

This module is designed in such a way that the participants get explicit idea regarding the Meteorology and its terms and concepts. Besides, we also wish that the participants will be enhance their official works. To achieve this objective, we have made the sessions based on the most important topics of Meteorology that are used in everyday life.

Key learning outcomes:

By the end of the course, delegates will have a knowledge and understanding of:

- Use an understanding of atmospheric processes to elucidate the practice of weather prediction.
- Access atmosphere science information from a variety of sources, evaluate the quality of this information, and compare this information with current models of meteorological processes, identifying areas of congruence and discrepancy.
- Use scientifically valid modes of inquiry, individually and collaboratively, to critically evaluate the hazards and risks posed by meteorological processes both to themselves and society as a whole, evaluate the efficacy of possible ethically robust responses to these risks, and effectively communicate the results of this analysis to their peers.
- Assess the contributions of meteorology to our evolving understanding of global change and sustainability while placing the development of meteorology in its historical and cultural context.

Disclaimer

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SESSION 1: INTRODUCTION TO METEOROLOGY

1.1 What is Meteorology

The branch of science concerned with the processes and phenomena of the atmosphere, especially as a means of forecasting the weather. Meteorology is the inter disciplinary scientific study of the atmosphere. The word 'meteor' is a variation on the Greek "meteoron", which is a term dealing with any objects that originate in the sky.

1.1.1 The History of Meteorology

Studies in the field stretch back millennia, though significant progress in meteorology did not occur until the 18th century. The 19th century saw modest progress in the field after observing networks formed across several countries. It wasn't until after the development of the computer in the latter half of the 20th century that significant breakthroughs in weather forecasting were achieved.

The development of meteorology is deeply connected to developments in science, math, and technology. The Greek philosopher Aristotle wrote the first major study of the atmosphere around 340 BCE. Many of Aristotle's ideas were incorrect, however, because he did not believe it was necessary to make scientific observations. A growing belief in the scientific method profoundly changed the study of meteorology in the 17th and 18th centuries.

Evangelista Torricelli, an Italian physicist, observed that changes in air pressure were connected to changes in weather. In 1643, Torricelli invented the barometer, to accurately measure the pressure of air. The barometer is still a key instrument in understanding and forecasting weather systems. In 1714, Daniel Fahrenheit, a German physicist, developed the mercury thermometer. These instruments made it possible to accurately measure two important atmospheric variables.

There was no way to quickly transfer weather data until the invention of the telegraph by American inventor Samuel Morse in the mid-1800s. Using this new technology, meteorological offices were able to share information and produce the first modern weather maps. These maps combined and displayed more complex sets of information such as isobars (lines of equal air pressure) and isotherms (lines of equal temperature). With these large-scale weather maps, meteorologists could examine a broader geographic picture of weather and make more accurate forecasts.

In the 1920s, a group of Norwegian meteorologists developed the concepts of air masses and fronts that are the building blocks of modern weather forecasting. Using basic laws of physics, these meteorologists discovered that huge cold and warm air masses move and meet in patterns that are the root of many weather systems. Military operations during World War I and World War II brought great advances to meteorology. The success of these operations was highly dependent on weather over vast regions of the globe. The military invested heavily in training, research, and new technologies to improve their understanding of weather. The most important of these new technologies was radar, which was developed to detect the presence, direction, and speed of aircraft and ships. Since the end of World War II, radar has been used and improved to detect the presence, direction, and speed of precipitation and wind patterns.

The technological developments of the 1950s and 1960s made it easier and faster for meteorologists to observe and predict weather systems on a massive scale. During the 1950s, computers created the first models of atmospheric conditions by running hundreds of data points through complex equations. These models were able to predict large-scale weather, such as the series of high- and low-pressure systems that circle our planet. TIROS I, the first meteorological satellite, provided the first accurate weather forecast from space in 1962. The success of TIROS I prompted the creation of more sophisticated satellites. Their ability to collect and transmit data with extreme accuracy and speed has made them indispensable to meteorologists. Advanced satellites and the computers that process their data are the primary tools used in meteorology today. (Sources: https://www.ausstormscience.com/meteorology-basics/)

1.2 General Meteorology

1.2.1. Fundamentals of meteorology

1.2.1.1 Composition of the Atmosphere

Table 1.1 shows the various gases present in a volume of air near the earth's surface. Notice that nitrogen (N2) occupies about 78 percent and oxygen (O2) about 21 percent of the total volume. If all the other gases are removed, these percentages for nitrogen and oxygen hold fairly constant up to an elevation of about 80 km (or 50 mi). At the surface, there is a balance between destruction (output) and production (input) of these gases. For example, nitrogen is removed from the atmosphere primarily by biological processes that involve soil bacteria. It is returned to the atmosphere mainly through the decaying of plant and animal matter. Oxygen, on the other hand, is removed from the atmosphere when organic matter decays and when oxygen combines with other substances, producing oxides. It is also taken from the atmosphere during breathing, as the lungs take in oxygen and release carbon dioxide. The addition of oxygen to the atmosphere occurs during photosynthesis, as plants, in the presence of sunlight, combine carbon dioxide and water to produce sugar and oxygen.

The concentration of the invisible gas water vapor, however, varies greatly from place to place, and from time to time. Close to the surface in warm, steamy, tropical locations, water vapor may account for up to 4 percent of the atmospheric gases, whereas in colder arctic areas, its concentration may dwindle to a mere fraction of a percent. Water vapor molecules are, of course, invisible. They become visible only when they transform into larger liquid or solid particles, such as cloud droplets and ice crystals. The changing of water vapor into liquid water is called condensation, whereas the process of liquid water becoming water vapor is called evaporation. In the lower atmosphere, water is everywhere. It is the only substance that exists as a gas, a liquid, and a solid at those temperatures and pressures normally found near the earth's surface (Fig. 1.1).

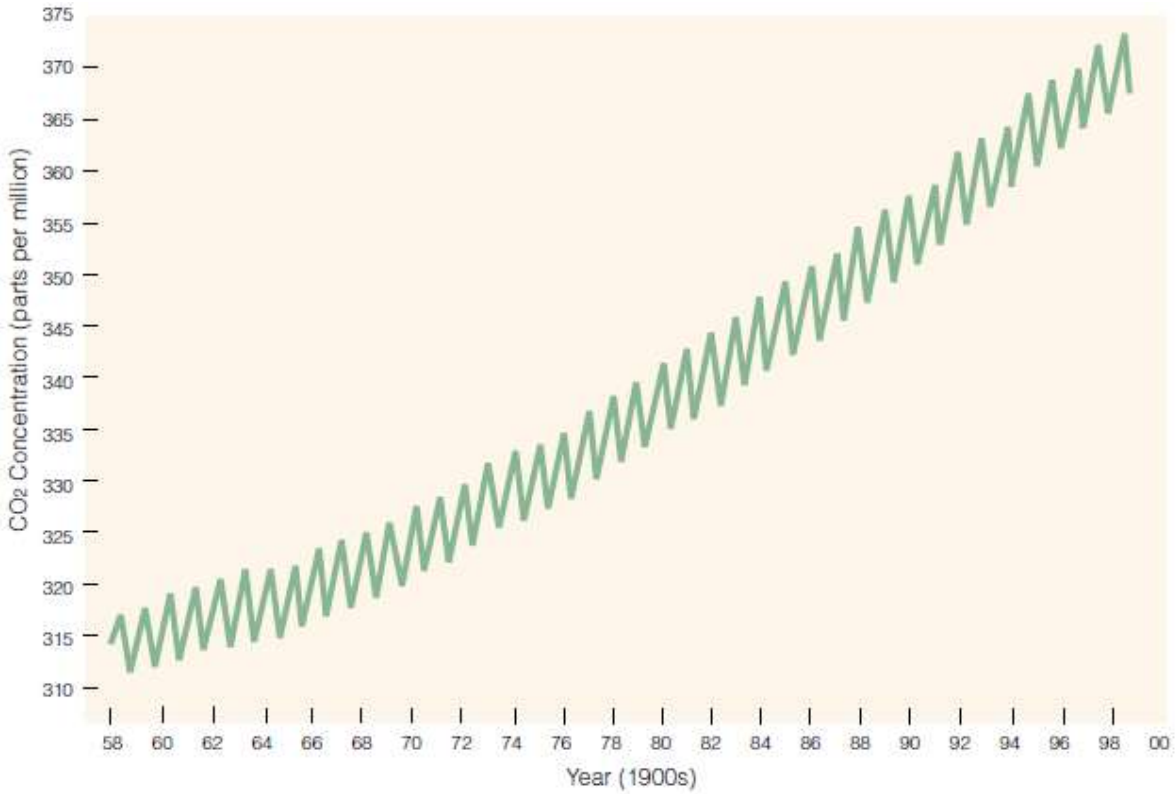
Water vapor is an extremely important gas in our atmosphere. Not only does it form into both liquid and solid cloud particles that grow in size and fall to earth as precipitation, but it also releases large amounts of heat- called latent heat -when it changes from vapor into liquid water or ice. Latent heat is an important source of atmospheric energy, especially for storms, such as thunderstorms and hurricanes. Moreover, water vapor is a potent greenhouse gas because it strongly absorbs a portion of the earth's outgoing radiant energy (somewhat like the glass of a greenhouse prevents the heat inside from escaping and mixing with the outside air). Thus, water vapor plays a significant role in the earth's heat energy balance.

Table 1.1: Composition of the Atmosphere Near the Earth's Surface

Permanent Gases			Variable Gases			
Gas	*Symbol*	*Percent (by Volume) Dry Air*	*Gas (and Particles)*	*Symbol*	*Percent (by Volume)*	*Parts per Million (ppm)**
Nitrogen	N_2	78.08	Water vapor	H_2O	0 to 4	
Oxygen	O_2	20.95	Carbon dioxide	CO_2	0.037	368*
Argon	Ar	0.93	Methane	CH_4	0.00017	1.7
Neon	Ne	0.0018	Nitrous oxide	N_2O	0.00003	0.3
Helium	He	0.0005	Ozone	O_3	0.000004	0.04†
Hydrogen	H_2	0.00006	Particles (dust, soot, etc.)		0.000001	0.01–0.15
Xenon	Xe	0.000009	Chlorofluorocarbons (CFCs)		0.00000002	0.0002

Fig. 1.1: The earth's atmosphere is a rich mixture of many gases, with clouds of condensed water vapor and ice crystals. Here, water evaporates from the ocean's surface. Rising air currents then transform the invisible water vapor into many billions of tiny liquid droplets that appear as puffy cumulus clouds. If the rising air in the cloud should extend to greater heights, where air temperatures are quite low, some of the liquid droplets would freeze into minute ice crystals.

Carbon dioxide (CO2), a natural component of the atmosphere, occupies a small (but important) percent of a volume of air, about 0.037 percent. Carbon dioxide enters the atmosphere mainly from the decay of vegetation, but it also comes from volcanic eruptions, the exhalations of animal life, from the burning of fossil fuels (such as coal, oil, and natural gas), and from deforestation. The removal of CO2 from the atmosphere takes place during photosynthesis, as plants consume CO2 to produce green matter. The CO2 is then stored in roots, branches, and leaves. The oceans act as a huge reservoir for CO2, as phytoplankton (tiny drifting plants) in surface water fix CO2 into organic tissues. Carbon dioxide that dissolves directly into surface water mixes downward and circulates through greater depths. Estimates are that the oceans hold more than 50 times the total atmospheric CO2 content.



Measurements of CO2 in parts per million (ppm) at Mauna Loa Observatory, Hawaii. Higher readings occur in winter when plants die and release CO2 to the atmosphere. Lower readings occur in summer when more abundant vegetation absorbs CO2 from the atmosphere.

Figure 1.2 reveals that the atmospheric concentration of CO2 has risen more than 15 percent since 1958, when it was first measured at Mauna Loa Observatory in Hawaii. This increase means that CO2 is entering the atmosphere at a greater rate than it is being removed. The increase appears to be due mainly to the burning of fossil fuels; however, deforestation also plays a role as cut timber, burned or left to rot, releases CO2 directly into the air, perhaps accounting for about 20 percent of the observed increase. Measurements of CO2 also come from ice cores. In Greenland and Antarctica, for example, tiny bubbles of air trapped within the ice sheets reveal that before the industrial revolution, CO2 levels were stable at about 280 parts per million (ppm). Since the early 1800s, however, CO2 levels have increased by as much as 25 percent. With CO2 levels presently increasing by about 0.4 percent annually (1.5 ppm/year), scientists now estimate that the concentration of CO2 will likely rise from its current value of about 368 ppm to a value near 500 ppm toward the end of this century.

Carbon dioxide is another important greenhouse gas because, like water vapor, it traps a portion of the earth's outgoing energy. Consequently, with everything else being equal, as the atmospheric concentration of CO2 increases, so should the average global surface air temperature. Most of the mathematical model experiments that predict future atmospheric conditions estimate that increasing levels of CO2 (and other greenhouse gases) will result in a global warming of surface air between 1°C and 3.5°C (about 2°F to 6°F) by the year 2100. Such warming could result in a variety of consequences, such as increasing precipitation in

certain areas and reducing it in others as the global air currents that guide the major storm systems across the earth begin to shift from their "normal" paths.

1.2.1.2 Vertical distribution of atmosphere

Earth's atmosphere is divided into layers or zones according to various distinguishing features. These divisions are for reference of thermal structure (lapse rates) or other significant features and are not intended to imply that these layers or zones are independent domains. Earth is surrounded by one atmosphere, not by a number of sub-atmospheres. The layers and zones are discussed under two separate classifications. One is the meteorological classification that defines zones according to their significance for the weather. The other is the electrical classification that defines zones according to electrical characteristics of gases of the atmosphere.

Meteorological Classification

In the meteorological classification (commencing with Earth's surface and proceeding upward) we have the troposphere, tropopause, stratosphere, stratopause, mesosphere, mesopause, thermosphere, and the exosphere. These classifications are based on temperature characteristics.

Troposphere: The troposphere is the layer of air enveloping Earth immediately above Earth's surface. It is approximately 5.5 miles (29,000 ft or 9 km) thick over the poles, about 7.5 miles (40,000 ft or 12.5 km) thick in the mid-latitudes, and about 11.5 miles (61,000 ft or 19 km) thick over the Equator. The figures for thickness are average figures; they change somewhat from day to day and from season to season. The troposphere is thicker in summer than in winter and is thicker during the day than during the night. Almost all weather occurs in the troposphere. However, some phenomena such as turbulence, cloudiness (caused by ice crystals), and the occasional severe thunderstorm top occur within the tropopause or stratosphere. The troposphere is composed of a mixture of several different gases. By volume, the composition of dry air in the troposphere is as follows: 78 percent nitrogen, 21 percent oxygen, nearly 1-percent argon, and about 0.03 percent carbon dioxide. In addition, it contains minute traces of other gases, such as helium, hydrogen, neon, krypton, and others. The air in the troposphere also contains a variable amount of water vapor. The maximum amount of water vapor that the air can hold depends on the temperature of the air and the pressure. The higher the temperature, the more water vapor it can hold at a given pressure. The air also contains variable amounts of impurities, such as dust, salt particles, soot, and chemicals. These impurities in the air are important because of their effect on visibility and the part they play in the condensation of water vapor. If the air were absolutely pure, there would be little condensation. These minute particles act as nuclei for the condensation of water vapor. Nuclei, which have an affinity for water vapor, are called hygroscopic nuclei. The temperature in the troposphere usually decreases with height, but there may be inversions for relatively thin layers at any level.

Tropopause: The tropopause is a transition layer between the troposphere and the stratosphere. It is not uniformly thick, and it is not continuous from the equator to the poles. In each hemisphere the existence of three distinct tropopauses is generally agreed upon- one in the subtropical latitudes, one in middle latitudes, and one in subpolar latitudes. They overlap each other where they meet. The tropopause is characterized by little or no change in temperature with increasing altitude. The composition of gases is about the same as that for the troposphere. However, water vapor is found only in very minute quantities at the tropopause and above it.

Stratosphere: The stratosphere directly overlies the tropopause and extends to about 30 miles (160,000 ft or 48 kilometers). Temperature varies little with height in the stratosphere through the first 30,000 feet (9,000 meters); however, in the upper portion the temperature increases approximately linearly to values nearly equal to surface temperatures. This increase in temperature through this zone is attributed to the presence of ozone that absorbs incoming ultraviolet radiation.

Stratopause: The stratopause is the top of the stratosphere. It is the zone marking another reversal with increasing altitude (temperature begins to decrease with height).

Mesosphere: The mesosphere is a layer approximately 20 miles (100,000 ft or 32 kilometers) thick directly overlaying the stratopause. The temperature decreases with height.

Mesopause: The mesopause is the thin boundary zone between the mesosphere and the thermosphere. It is marked by a reversal of temperatures, i.e., temperature again increases with altitude.

Thermosphere: The thermosphere, a second region in which the temperature increases with height, extends from the mesopause to the exosphere.

Exosphere: The very outer limit of Earth's atmosphere is regarded as the exosphere. It is the zone in which gas atoms are so widely spaced they rarely collide with one another and have individual orbits around Earth.

Electrical Classification

The primary concern with the electrical classification is the effect on communications and radar. The electrical classification outlines three zones—the troposphere, the ozonosphere, and the ionosphere.

Troposphere: The troposphere is important to electrical transmissions because of the immense changes in the density of the atmosphere that occur in this layer. These density changes, caused by differences in heat and moisture, affect the electronic emissions that travel through or in the troposphere. Electrical waves can be bent or refracted when they pass through these different layers and the range and area of communications may be seriously affected.

Ozonsophere: This layer is nearly coincident with the stratosphere. As was discussed earlier in this section, the ozone is found in this zone. Ozone is responsible for the increase in temperature with height in the stratosphere.

Inosphere: The ionosphere extends from about 40 miles (200,000 ft or 64 kilometers) to an indefinite height. Ionization of air molecules in this zone provides conditions that are favorable

for radio propagation. This is because radio waves are sent outward to the ionosphere and the ionized particles reflect the radio waves back to Earth.

Circulation of Water in the Atmosphere

Within the atmosphere, there is an unending circulation of water. Since the oceans occupy over 70 percent of the earth's surface, we can think of this circulation as beginning over the ocean. Here, the sun's energy transforms enormous quantities of liquid water into water vapor in a process called evaporation. Winds then transport the moist air to other regions, where the water vapor changes back into liquid, forming clouds, in a process called condensation. Under certain conditions, the liquid (or solid) cloud particles may grow in size and fall to the surface as precipitation—rain, snow, or hail. If the precipitation falls into an ocean, the water is ready to begin its cycle again. If, on the other hand, the precipitation falls on a continent, a great deal of the water returns to the ocean in a complex journey. This cycle of moving and transforming water molecules from liquid to vapor and back to liquid again is called the hydrologic (water) cycle. In the most simplistic form of this cycle, water molecules travel from ocean to atmosphere to land and then back to the ocean.

For example, before falling rain ever reaches the ground, a portion of it evaporates back into the air. Some of the precipitation may be intercepted by vegetation, where it evaporates or drips to the ground long after a storm has ended. Once on the surface, a portion of the water soaks into the ground by percolating downward through small openings in the soil and rock, forming groundwater that can be tapped by wells. What does not soak in collects in puddles of standing water or runs off into streams and rivers, which find their way back to the ocean. Even the underground water moves slowly and eventually surfaces, only to evaporate or be carried seaward by rivers. Over land, a considerable amount of vapor is added to the atmosphere through evaporation from the soil, lakes, and streams. Even plants give up moisture by a process called transpiration. The water absorbed by a plant's root system moves upward through the stem and emerges from the plant through numerous small openings on the underside of the leaf. In all, evaporation and transpiration from continental areas amount to only about 15 percent of the nearly 1.5 billion gallons of water vapor that annually evaporate into the atmosphere; the remaining 85 percent evaporates from the oceans. The total mass of water vapor stored in the atmosphere at any moment adds up to only a little over a week's supply of the world's precipitation. Since this amount varies only slightly from day to day, the hydrologic cycle is exceedingly efficient in circulating water in the atmosphere.

Evaporation, Condensation, and Saturation

To obtain a slightly different picture of water in the atmosphere, suppose we examine water in a dish similar to the one shown in Fig. 1.3. If we were able to magnify the surface water about a billion times, we would see water molecules fairly close together, jiggling, bouncing, and moving about. We would also see that the molecules are not all moving at the same speed—some are moving much faster than others. Recall from Chapter 2 that the temperature of the water is a measure of the average speed of its molecules. At the surface, molecules with enough speed (and traveling in the right direction) would occasionally break away from the

liquid surface and enter into the air above. These molecules, changing from the liquid state into the vapor state, are evaporating. While some water molecules are leaving the liquid, others are returning. Those returning are condensing as they are changing from a vapor state to a liquid state. When a cover is placed over the dish (Fig. 1.4), after a while the total number of molecules escaping from the liquid (evaporating) would be balanced by the number returning (condensing). When this condition exists, the air is said to be saturated with water vapor. For every molecule that evaporates, one must condense, and no net loss of liquid or vapor molecules results. If we remove the cover and blow across the top of the water, some of the vapor molecules already in the air above would be blown away, creating a difference between the actual number of vapor molecules and the total number required for saturation. This would help prevent saturation from occurring and would allow for a greater amount of evaporation. Wind, therefore, enhances evaporation. The temperature of the water also influences evaporation. All else being equal, warm water will evaporate more readily than cool water. The reason for this phenomenon is that, when heated, the water molecules will speed up. At higher temperatures, a greater fraction of the molecules has sufficient speed to break through the surface tension of the water and zip off into the air above. Consequently, the warmer the water, the greater the rate of evaporation. If we could examine the air above the water in either Fig. 1.3 or Fig. 1.4, we would observe the water vapor molecules freely darting about and bumping into each other as well as neighboring molecules of oxygen and nitrogen. We would also observe that mixed in with all of the air molecules are microscopic bits of dust, smoke, and salt from ocean spray. Since many of these serve as surfaces on which water vapor may condense, they are called condensation nuclei. In the warm air above the water, fast-moving vapor molecules strike the nuclei with such impact that they simply bounce away. However, if the air is chilled, the molecules move more slowly and are more apt to stick and condense to the nuclei. When many billions of these vapor molecules condense onto the nuclei, tiny liquid cloud droplets form. We can see then that condensation is more likely to happen as the air cools and the speed of the vapor molecules decreases. As the air temperature increases, condensation is less likely because most of the molecules have sufficient speed (sufficient energy) to remain as a vapor. As we will see in this and other chapters, condensation occurs primarily when the air is cooled. Even though condensation is more likely to occur when the air cools, it is important to note that no matter how cold the air becomes, there will always be a few molecules with sufficient speed (sufficient energy) to remain as a vapor. It should be apparent, then, that with the same number of water vapor molecules in the air, saturation is more likely to occur in cool air than in warm air. This idea often leads to the statement that "warm air can hold more water vapor molecules before becoming saturated than can cold air" or, simply, "warm air has a greater capacity for water vapor than does cold air." At this point, it is important to realize that although these statements are correct, the use of such words as "hold" and "capacity" are misleading when describing water vapor content, as air does not really "hold" water vapor in the sense of making "room" for it.

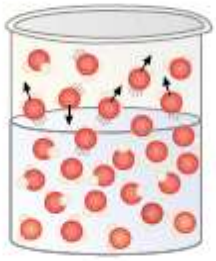


Fig. 1.3: Water molecules at the surface of the water are evaporating (changing from liquid into vapor) and condensing (changing from vapor into liquid). Since more molecules are evaporating than condensing, net evaporation is occurring. (For clarity, only water molecules are illustrated.)

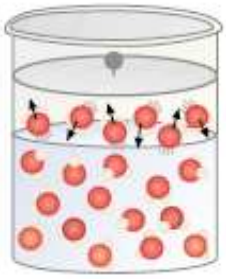


Fig. 1.4: When the number of water molecules escaping from the liquid (evaporating) balances those returning (condensing), the air above the liquid is saturated with water vapor. (For clarity, only water molecules are illustrated.)

1.2.1.3 Relative Humidity

Relative humidity is also defined as the ratio (expressed in percent) of the observed vapor pressure to that required for saturation at the same temperature and pressure. Relative humidity shows the degree of saturation, but it gives no clue to the actual amount of water vapor in the air. Thus, other expressions of humidity are useful.

Absolute Humidity: The mass of water vapor present per unit volume of space, usually expressed in grams per cubic meter, is known as absolute humidity. It may be thought of as the density of the water vapor.

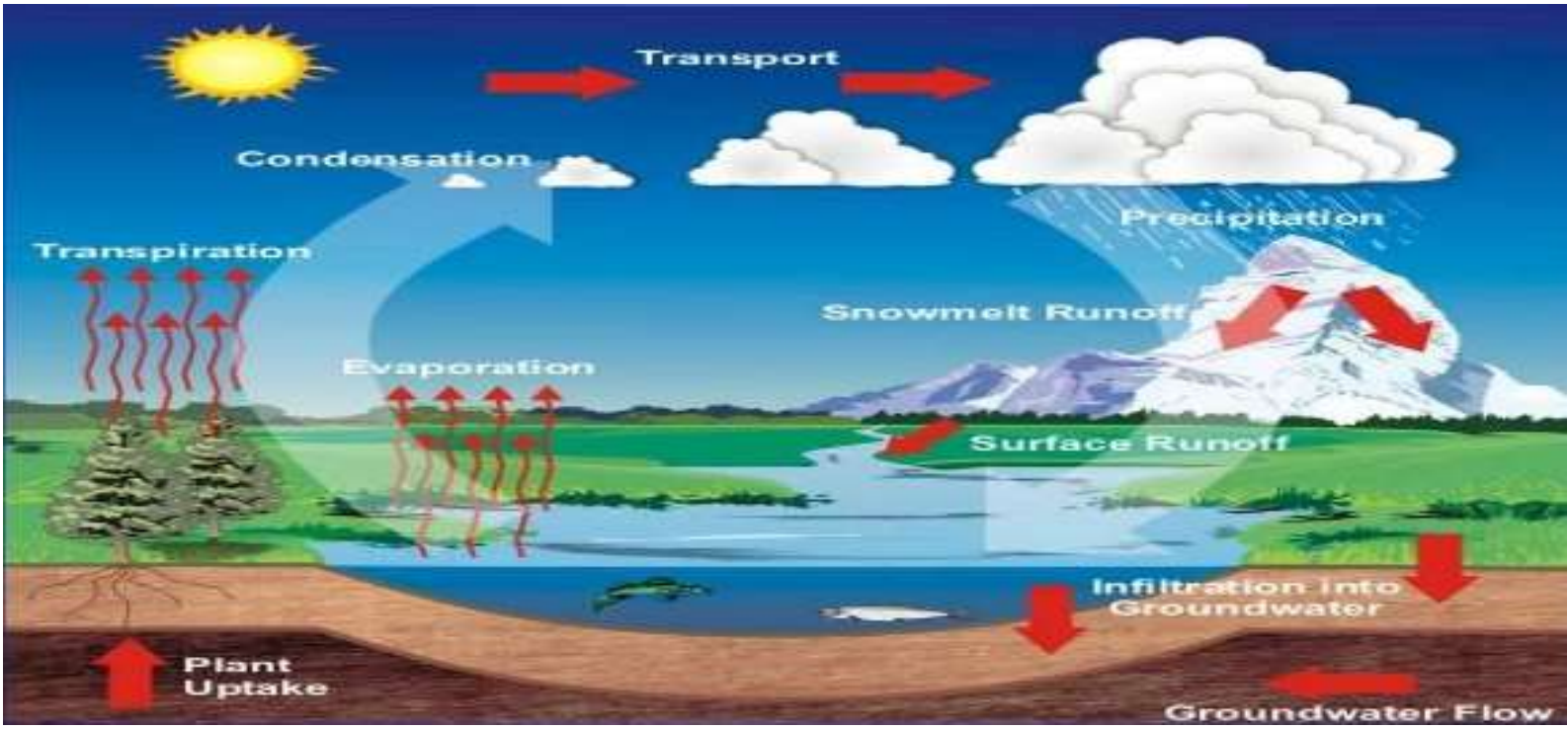
Specific Humidity: Humidity may be expressed as the mass of water vapor contained in a unit mass of air (dry air plus the water vapor). It can also be expressed as the ratio of the density of the water vapor to the density of the air (mixture of dry air and water vapor). This is called the specific humidity and is expressed in grams per gram or in grams per kilogram. This value depends upon the measurement of mass, and mass does not change with temperature and pressure. The specific humidity of a parcel of air remains constant unless water vapor is added to or taken from the parcel. For this reason, air that is unsaturated may move from place to place or from level to level, and its specific humidity remains the same as long as no water vapor is added or removed. However, if the air is saturated and cooled, some of the water vapor must condense; consequently, the specific humidity (which reflects only the water vapor) decreases. If saturated air is heated; its specific humidity remains unchanged unless water vapor is added to it. In this case, the specific humidity increases. The maximum specific humidity that a parcel can have occurs at saturation and depends upon both the temperature and the pressure. Since warm air can hold more water vapor than cold air at constant pressure, the saturation specific humidity at high temperatures is greater than at low temperatures. Also, since moist air is less dense than dry air at constant temperature, a parcel of air has a greater specific humidity at saturation if the pressure is low than when the pressure is high.

Mixing Ratio: The mixing ratio is defined as the ratio of the mass of water vapor to the mass of dry air and is expressed in grams per gram or in grams per kilogram. It differs from specific humidity only in that it is related to the mass of dry air instead of to the total dry air plus water vapor. It is very nearly equal numerically to specific humidity, but it is always slightly greater. The mixing ratio has the same characteristic properties as the specific humidity. It is conservative (values do not change) for atmospheric processes involving a change in temperature. It is non conservative for changes involving a gain or loss of water vapor. Previously it was learned that air at any given temperature can hold only a certain amount of water vapor before it is saturated. The total amount of vapor that air can hold at any given temperature, by weight relationship, is referred to as the saturation mixing ratio. It is useful to note that the following relationship exists between mixing ratio and relative humidity. Relative humidity is equal to the mixing ratio divided by the saturation mixing ratio, multiplied by 100. If any two of the three components in this relationship are known, the third may be determined by simple mathematics.

Dew Point: The dew point is the temperature that air must be cooled, at constant pressure and constant water vapor content, in order for saturation to occur. The dew point is a conservative and very useful element. When atmospheric pressure stays constant, the dew point reflects increases and decreases in moisture in the air. It also shows at a glance, under the same conditions, how much cooling of the air is required to condense moisture from the air.

2.1 Evaporation

Measurement of evaporation from free water surface and soil and transpiration from vegetation are of great importance in agricultural and hydro meteorological studies. The rate of evaporation is defined as the amount of water lost by evaporation from a unit area of surface in unit time. This is equivalent to a certain depth of liquid water lost per unit time from the whole area. Evaporation is measured by observing the change of level of free water surface in a pan or a tank.



Nature of Process

By evaporation water in liquid state is changed to vapor state. This change occurs when some molecules in the water mass have attained enough kinetic energy to erect themselves from the water surface. Escaping molecules of course are attracted by other molecules which tend to hold them together with in the water. Only, those molecules possessing gather kinetic energy then the average within the liquid escape from the surface. The temperature of the liquid is lower by such escape, so evaporation results in cooling. Molecules may leave a solid surface in the same way. This change without passing through the usual intermediate liquid state is called sublimation.

The motion of the molecules through the water surface produces a pressure. This pressure of the aqueous vapor is called vapor pressure. More correctly it is only the partial pressure of

water vapor in the atmosphere, because in a mixture of gases each gas exerts a partial pressure which is independent of the other gases in the mixture.
Escaping molecules collide with those in the air and some of the former will drop back into the water. When the number of molecules that escape equals the number of those that falls back into the water equilibrium is reached between the pressures exerted by the escaping molecule and the pressure of the surrounding atmosphere. Further, some of the molecules in the gaseous phase have kinetic energy sufficient to cause them to penetrate in the liquid and others will condense from a vapor to a liquid state. Thus, evaporation from and condensation on the liquid surface are continuous processes. Evaporation is faster than condensation if the space above the water surface is not saturated.

Errors affecting evaporation: The rate of evaporation depends on vapor pressure of the body of water and that of the air. This vapor pressure depends on:

i) Total radiation and terrestrial.
ii) Temperature, both of the air and the evaporating surface.
iii) Wind speed at the surface.
iv) Atmospheric pressure.
v) Nature of surface.
vi) Amount of moisture in the surface available for evaporation.

1) Temperature: The vapor pressure of a body of water increases with temperature, because the kinetic energy of the water molecules is raised with creasing temperature. Since vapor pressure is proportional to the vapor pressure difference between the water and the air, equal temperature increases may not increase the rate of evaporation. For evaporation to continue, heat must be applied to the water as it is cooled by evaporation. Otherwise, while air and water temperature become equal, evaporation ceases.

2) Wind: Water molecules escaping from the water surface collide with others already present in the air; wind is effective in remaining water molecules in the air and bringing it in capable of holding more water vapor. When the wind velocity is great enough to remove all of the water molecules escaping from the water surface a further increase in velocity will not increase evaporation appreciably.

The effect of wind on evaporation may be more pronounced over large bodies of water than over small areas. In case of pans, a slight increase in wind may just as it appears.

United States Class A Pan: USA Class A pan is used in BMD for measurement of evaporation. The pan is of cylindrical design, 25cm. deep and 120.7cm. in diameter. It is constructed of galvanized iron and left unpainted. The pan is filled with water to 5cm. below the rim.

The water level is measured by means of either a hook gauge or a fixed-point gauge. The hook gauge consists of a moveable scale and vernier fitted with a hook, the point of which indicates when the gauge is correctly set to touch the water surface. A stilling well, about 10cm. across

30cm deep with a small hole at the bottom, breaks any ripples in the main part of the tank, and serves as a support for the hook gauge during an observation. The pan is refilled whenever the water level drops by more than 2.5cm.

The fixed gauge consists of a pointed brass rod fixed in a stilling well so that its tip is located six to seven centimeters below the rim of the pan. A calibrated container issued to add or remove water ateach observation to return the water level to the fixed point.

Evaporograph

Principle of operation: The principle of operation consists of water evaporating out of the reservoir through a round disc of filter paper. As the water evaporates a float located in the reservoir moves downward. This float is mechanically attached to the pen arm of the recorder. Used outdoors, the instrument is exposed to the open bodies of water. It may be placed on a wooden frame not much above the ground level.

Pre-operational setting:

1. Install a clean filter paper disc on the evaporation surface. The filter paper keeps the cloth wick. The filter paper is installed by removing the filter ring.
2. Remove the clock drum.
3. Wind the clock approximately seven turns. Do not force winding key.
4. Remove the chart clip from the side of the clock drum. Select the chart and fill in the station name and date the chart is started. Place the right- hand margin of the chart in line with the right side of the chart clip notch. Wrap the chart around the drum. The left-hand edge of the chart should overlap the right-hand edge and horizontal scale line on the two edges of the chart should coincide. Place the clip in position.
5. Place the clock drum on the gear post and secure it with the knurled nut.
6. Set the pen on the chart with the help of the pen lifter on the base of the instrument.

Maintenance:

1. The instrument should be routinely maintained preventing dust and dirt buildup. When a chart change is required, simply brush the dust and dirt from the instrument.
2. At routine intervals, 3 or 6 months, clean all pivot points with a small brush apply a light coat of instrument oil.

3. Inspect the filter paper each time the chart changed, or water is added to the cistern. If the filter paper is soiled replace it with a new one. Never operate the instrument without the filter paper installed.
4. Use only distilled water in the cistern as otherwise mineral deposits might from in water tubing and vitiate the performance of the instrument.

2.2 Humidity Measuring Instrument

- Relative humidity: The ratios in per cent of the observed vapour pressure to the saturation vapour pressure with respect to water at the same temperature and pressure; $H= (e/e_w) \times 100$
- Vapour pressure: The partial pressure of water vapour in air.
- Saturation vapour pressures: Vapour pressures in air in equilibrium with the surface of water and ice, respectively

Hygrometer:

04 types hygrometer are used for relative humidity measurement. They are-

Hygrometer
- Phychrometer
- Hair Hygrograph
- Elecelectric Hygrometertric Hygrometer
- Chilled- Mirror Dew Point Hygrometer

Mainly Phychrometer hygrometer & Hair hygrograph are used in Bangladesh Meteorological Department for humidity measurement.

Principle of a Psychrometer:

When water or ice covers the bulb of wet bulb thermometer; latent heat is removing from the surface of the bulb as the water evaporates and the temperature of the bulb becomes lower than the air temperature (dry bulb). At a lower humidity water evaporates more actively; so that the temperature of wet bulb thermometer will be lower rapidly. The phychrometer measures humidity by difference between the dry bulb and wet bulb temperature with respect to hygrometric table.

Type of Psychrometer

There are two types of portable phychrometer is use in BMD. They are-

1. Assuman phychrometer: An Assuman aspirated phychrometer consists essentially of a Perspex duct in which are supported two Assuman type thermometers and through which a current of air is drawn by means of a fan driving by a powerful clockwork motor. The thermometers are supported by a highly polished double walled radiation shield which can be removing when it is necessary to moisten the wet bulb.

Methods observation by Assuman phychrometer:

i. Moisten the wet bulb thermometer with the help of an injector or dropper. To do this remove the radiation shield, then fill the injector with distilled water and pass the tube up over the wet bulb.
ii. Wind the clockwork motor fully.
iii. Place the instrument in the position on the post.
iv. Start reading the wet bulb at intervals of about 10 sec, after 1.5 min. of the working of the fan from start. When the wet-bulb temperature readings have become steady, then both the dry bulb and wet bulb temperature should be read. Take readings to the nearest 0.1˚C.

2. Whirling Phychrometer: This is a small portable type phychrometer and consists of two mercury-in-glass thermometers mounted in a boxwood frame which is provided with a handle and spindle, by means of which the frame and thermometer can be rotated quickly about a horizontal axis. In this way the thermometer bulbs pass rapidly through air and they are ventilated as in an as aspirated phychrometer.

Exposure of phychrometer:

For measurement of air temperature and humidity, the phychrometer should be hold above 4ft from the ground in an open place and the instrument at the wind wards side. It may be held in the hand keeping away from the body and to the wind wards.

Methods observation by Whirling Phychrometer:

i) Stand in shade facing the wind and rotate the frame rapidly, at the rate of at least 3revolutions per second, for about a minute and then take the reading rapidly.

ii) Take several readings each time after rotating the frame rapidly. Successive readings should not differ by more than 0.2 ˚C. Take readings to the nearest 0.1˚C.

Maintenance

Routine maintenance

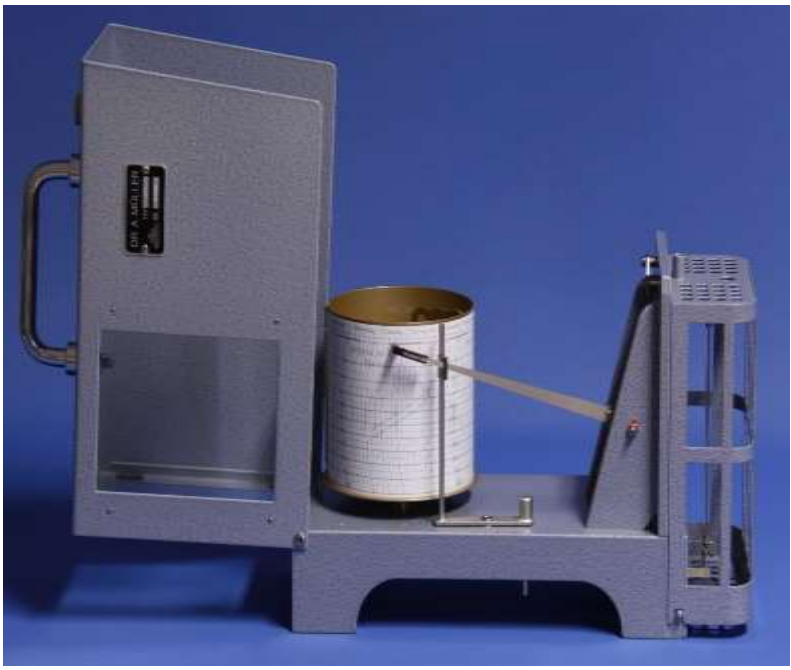
Make sure the aspiration fan in running properly at every observation.

Periodic maintenance

a) If the wet sleeve becomes dirty, replace it with a new one. In Japan, the sleeve is replaced twice a month. However, the sleeve must be replaced more frequently if it is placed in the environment with much dirt or in the sea breeze zone.

b) Wash off deposits on the wet bulb when the wet sleeve is replaced.

Principle of Hair hygrometer: The hair hygrometer uses the characteristic of the hair that its length expands or shrinks response to the relative humidity. The dimensions of various organic materials vary with their moisture content. A humidity change takes an effect on the moisture content in such materials. The length of human hair from which liquid are removed increases by 2 to 2.5% when relative humidity changes by 0 to 100%. Different types of human hair show different changes in length. However, there is still a relationship between the length of hair and relative humidity. The hair hygrograph is a hair hygrometer to which a clock-driven drum is installed to record humidity not a recording chart.



Precautions for using the hair hygrograph:

1) Before taking a reading of the hair hygrograph, gently tap the hygrometer to remove any mechanical tension added to the hair bundle.
2) At every measurement with the hair hygrograph, the reading should be compared with the humidity measured with the aspirated psychrometer at the same time. The difference of the humidity between them is used as a correction value.
3) Time marks as well as the degree of clock accuracy should be recorded on the chart.
    a. When making a time mark on the recording chart by moving the pen, take care to move the pen arm downward. Moving the pen arm in the pen arm in the opposite direction (upward) makes the hair bundle to expand, causing the hygrograph to become defective.
    b. To determine the humidity from the recording chart, read the indication on the record then correct it with correction.

Maintenance

Routine maintenance

a) Clean the dusty hair bundle with dust or smoke with a soft brush.
b) If the hair bundle is extremely dirty or it has been used for several months, clean it with a painting brush soaked with distilled warm water by gently touching the bundle.
c) Do not touch the hair bundle directly. Each of the parts of the hair hygrometer operates under slight forces. When cleaning the parts, be sure to treat them gently.

Periodic maintenance

a) Clean the hair bundle with a feather brush; wash it with a painting brush.
b) Make offset adjustments by comparing the reading of the hair hygrometer with that of the aspirated psychrometer.

Elecelectric Hygrometertric Hygrometer

Chilled- Mirror Dew Point Hygrometer

2.3 Measurement of soil moisture

There are several methods currently in use for measurement of soil moisture; only the (i) direct gravimetric method and (ii) neutron scattering method have been recommended by WMO.

Direct Gravimetric sampling of soil moisture: In this method the water contain of the soil is measured by oven drying samples at 100-105^{0}C until there is no further change of the weight.

Required Instrument for this method:

(i)Thermostatically controlled oven; (ii) Balance; (iii) Auger; (iv) porcelain cup (05Nos.) and (v) Desiccators.

Collection of samples: Soil samples are obtained with the aid of auger. It consists of a sampling cylinder with a spiral cutting edge fitted into the soil by means of the rod and the handle to the required depth which is marked on the rod. The auger is then withdrawn, and surplus soil removed with a special cleaning tool. Some 20 to 40grams of soil sample are taken from the lowest third of sampling cylinder and placed in special, numbered container with tightly fitting lid. A special carrying case is used to transport the sample to the laboratory.

Measurement of soil moisture: The soil samples are then taken to the laboratory and after the exterior of the containers have been cleaned, are immediately weighted to an accuracy of 0.05 gram. The soil samples are then dried in a thermostatically controlled oven at a temperature of 100-105˚C. The time needed for drying depends on both the moisture content and the type of the soil. The drying times required usually vary between 16 and 24 hours.

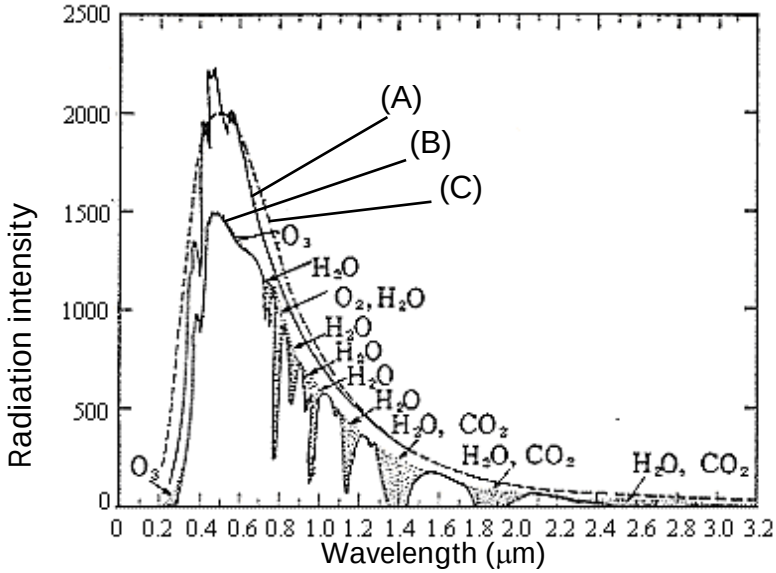
After the first drying, the samples are weighted again and then returned to the drying oven for a further two hours and then again weighted. If the difference between the two weights after each drying does not exceed 0.1gram, the soil is emptied, and the container weighted.

Precaution to be taken:

a) The observation or sampling site should be large enough to allow for repeated withdrawal of samples without rendering the site unrepresentative of local agricultural practice.
b) Considerable attention should be given to the sampling procedure because of the great horizontal and vertical variability of the soil moisture content. This is of particular importance when routine measurements are made to obtain data representative of agricultural conditions.
c) Dried soil samples are very hygroscopic and should not be exposed to the air whilst awaiting weighting. Desiccators should be used at this phase of the wor.

2.4 Measurement of Solar Radiation

Everything in nature emits electromagnetic energy, and solar radiation is energy emitted by the sun. The energy of extraterrestrial solar radiation is distributed over a wide continuous spectrum ranging from ultraviolet to infrared rays. In this spectrum, solar radiation in short wavelengths (0.29 to 3.0 μm) accounts for about 97% of the total energy. The below figure shows the spectrum distribution of solar radiation.



(A) Extraterrestrial solar radiation
(B) Direct solar radiation on the earth's surface
(C) Blackbody radiation at 5,900 K

(Shaded areas indicate absorption by the atmosphere.)

Spectrum distribution of solar radiation

Solar radiation is partly absorbed, scattered and reflected by molecules, aerosols, water vapor and clouds as it passes through the atmosphere. The direct solar beam arriving directly at the earth's surface is called direct solar radiation. The total amount of solar radiation falling on a horizontal surface (i.e. the direct solar beam plus diffuse solar radiation on a horizontal surface) is referred as global solar radiation.

Direct solar radiation is observed from sun rise to sunset, while global solar radiation is observed in the twilight before sunrise and after sunset, despite its diminished intensity at these times.

Actinograph/ Pyranograph:

The Actinograph is an instrument for measuring direct rays of the sun combined with rays scatted in the air and reaching the ground that is light in the air. However, as both rays direct from the sun and in the air pass through the air and further through the glass globe, those which reach the bimetal have a wavelength ranging from 03μ to 3.0μ. When Rays in this range of the wavelength reach the bimetal, they are reflected from the white piece while absorbed by its black piece. The black piece rises in temperature owing to the absorption of heat, resulting in the difference in the bend between the black and white pieces. This difference is recorded by a self-registering pen.

Installation of Pyranograph

1. The instrument should be installed in such a place as not to be shadowed from sunshine till sunset. The neighbourhood should be free from objects which will give strong reflected rays to the instrument.
2. The longer axis of the bimetal shall be placed exactly in a north-south direction, with the side window facing east.
3. The instrument should be properly levelled with the help of the attached level by adjusting two set screws.
4. The chart should be changed about one hour and a half after the sunset.

Maintenance and care of Pyranograph:

1. When the cover is removed, the outside and inside surfaces of the glass globe should be carefully wiped.
2. When the sun is not shining on the tip of the pen, the pen should be placed a few mm above the bottom of the chart.
3. The glass globe should be carefully handled to prevent damaging it.
4. Care should be taken to see that the instrument does not get any jerk.
5. Silica gel should be used as desiccant.
6. The levers from the bimetal to pen's rotary shaft shall not be moved as they are related to the adjustment of magnification
7. If moisture condenses inside the glass globe or on the bimetal surface to make them dim, the valve should be repaired.
8. In the following cases the instrument shall be re-examined:
    a) When the bimetal is changed or recoated.
    b) When the glass globe is exchanged.
    c) When the lever fitted parts undergo a change due to adjustment, etc.
9. During bad weather, the instrument should be removed in doors.

2.5 Pilot Balloon Theodolite

Main parts of a Pilot Balloon Theodolite:

1. Main Telescope.	16. Horizontal buble.
2. Graticule illumination control knob.	17. Battery box.
3. Auxiliary telescope.	18. Lamp switch.
4. Open sight.	19. Graticule illumination lamp.
5. Eyepiece.	20. Circular bubble.
6. Azimuth tangent screw release lever.	21. Rear circle reading window.
7. Azimuth reading drum.	22. Plumb bob stirrup.
8. Circle reading lens.	23. Spring plate clamping screw.
9. Azimuth lower tangent screw.	24. Azimuth lower clamping screw.
10. Foot screws	25. Circle illumination lamp.
11. Spring plate.	26. Graticule adjustment screw.
12. Stopwatch hook.	27. Mirror lever.
13. Elevation tangent screw release lever.	28. Horizontal bubble adjusting nuts.
14. Elevation reading drum.	29. Object glass adjustment screw.
15. From circle reading window.	

Introduction: The theodolite basically consists of-

a) A telescope mounted so that it can rotate in both azimuth and elevation.
b) Divided circles to measure the angles through which it is rotated.

A characteristic feature of pilot balloon theodolites is the right-angled telescope, which results in the position of the eyepiece remaining constant irrespective of the angle of elevation of the telescope.

Description: The Pilot Balloon Theodolite incorporates a main telescope in which the line of sight is turned through 90^0 via a prism in the body of the telescope. To enable the balloon to be brought quickly into the field of view of the main telescope, a low power auxiliary telescope with a wide field of view is provide. It views through the same eyepiece by turning a lever which swings a provide mirror into the optical path.

For night observations, two lamps are provided, one of which illuminates the scales and reading drums, and is so positioned that is also illuminates a stopwatch suspended from the hook. The other illuminates the graticule in the eyepiece, the intensity of the graticule illumination being controlled by the control knob. The graticule is divided to 1/1000 radian for use when tail observations are made. Both lamps are supplied from batteries in the detachable battery box and are operated independently by the switches.

The azimuth and elevation circles are arranged horizontally with the elevation circle uppermost. Both circles are read through the same window is situated on which is engraved a common index line. This window is situated at the front of the instrument below the telescope eyepiece. It is provided with a reading lens which can be swung downwards if not required. Another window at the rear of the instrument enables readings to be taken by the second observer. Both circles are graduated at 0.5^0intervls, the azimuth circle from 0^0 to 360^0, the elevation circle from 0^0 to 90^0 on either side of the vertical.

The instrument is directed by rotation of the two reading drums which are graduated at 0.1^0 intervals. They carry short handles to enable the instrument to be traversed and elevated rapidly. The elevation reading drum is on the left of the scale window with the tangent screw release lever immediately below. The azimuth reading drum is on the right, with its tangent screw release lever above.

- To avoid confusion, the elevation circle is filled in red, the azimuth circle in white.
- To enable the instrument to be accurately leveled, a horizontal is provided. For rough leveling the circular bubble is provided. The leveling is done by adjusting the foot screws.

2.6 Measurement of Atmospheric Pressure

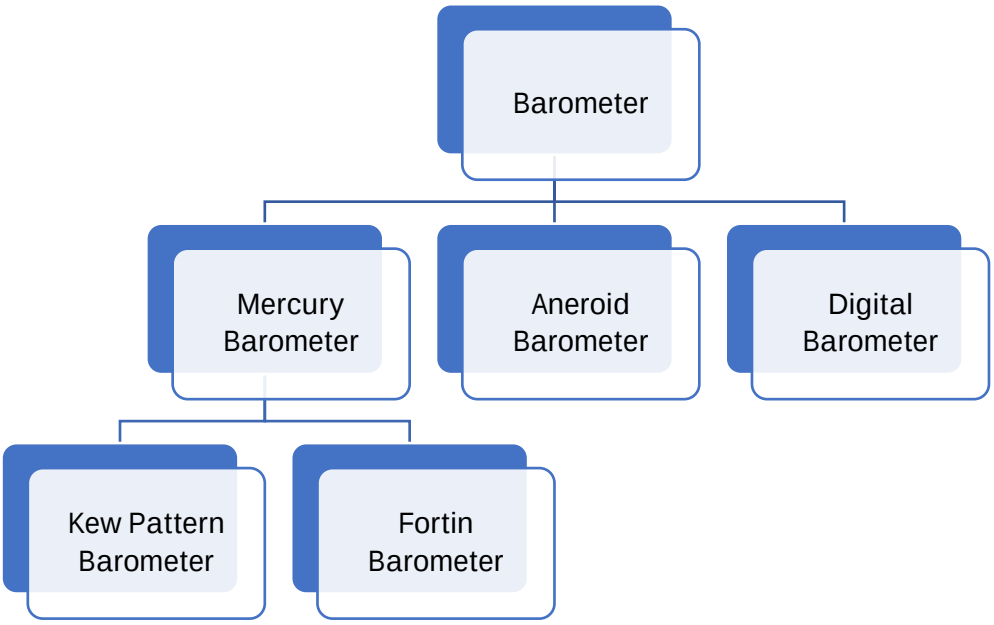
Atmospheric pressure: The pressure of the atmosphere at a given surface is the force per unit area exerted by virtue of its weight and is thus equal to the weight of vertical column of the air of unit area above the surface in question extending outer limit of the atmosphere.

Units of atmospheric pressure: Near the earth's surface the air exerts a pressure of about 10^5 Newton on each square meter of surface. This pressure is equal to 10^5 pascals (= one bar). Small variations in pressure occur during the day, and it is therefore, desirable to use a smaller unit, so that these changes may be reported. The unit used for Meteorological measurement is one hundred Pascals. It is called the hectopascal (= one milibar).

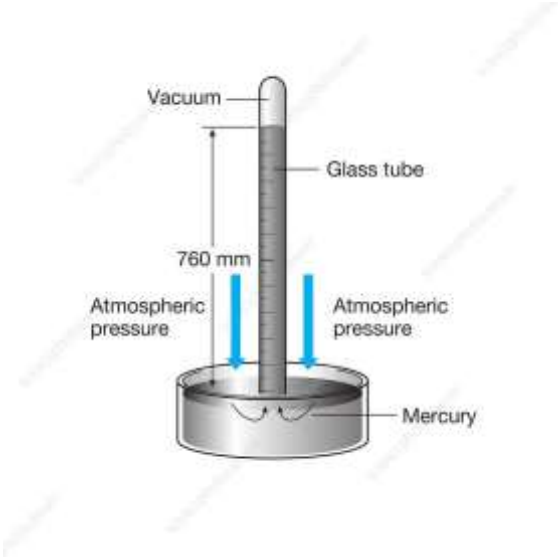
i.e. 10^5pascals = one bar= 1000hPa.

A hectopascal is, therefore, equivalent to a force of 100 Newton on each square meter of surface in contact with the air.

Instrument for Measurement Atmospheric pressure: The Atmospheric pressure is measured by means of Barometer. Three most common types of Barometer are:



Principle of the Mercury Barometer: The basic principle of the mercury barometer is that the pressure of the atmosphere is balanced against the weight of the mercury column of the mercury. The mercury column in the tube is supported by the pressure of the air on the surface of the mercury in the cistern.



Function of the mercury Barometer: As the mercury in the barometer tube rises or falls the mercury level in the cistern changes in the opposite direction and unless this change is taken into account the readings of the height of the mercury column in the tube will not represent the actual pressure of the air. For the meteorological purposes the length of the mercury column is measured on a scale, graduated in units of pressure i.e.in mb or hectopascal.

Why mercury is used in Barometer:

Mercury is used as the barometric fluid for the following reason:

a) Its large relative density makes the column of convenient length (Density of mercury is 13.5951g cm^{3}).
b) Its vapor pressure so small at ordinary working temperature.
c) It is easily cleaned and purified.
d) It does not wet the wall of tube, So, that the meniscus is convex upward.

This make it easy so measure its position with accurately.

Main parts of Barometer:

The Barometer has got five main parts. They are-

i. Mercury tube.
ii. Cistern.
iii. Brass case.
iv. Main scale & vernier scale.
v. Attached thermometer.

i) Mercury tube: Tube of KP barometer is not uniform bore. The bore is about 1.6mm for most of the way, but in the visible part at the top of the tube is about 8mm. There is an air trap inside the tube, which consists of an inverted pipette arrangement; any air bubbles which may enter at the bottom of the tube tend to keep to the sides and will not therefore pass to the vacuum at the top of the tube .Instead, they will be collected in the shoulders of the trap. Tube of fortin barometer is uniform bore.

ii) Cistern: Cistern is of steel and screws into cistern top against a leather washer. Two small vent holes in the cistern top allow the passage of the air the space above the mercury. Usually, the cistern has an internal diameter of 50mm and depth 35mm.

iii) Brass case: For protection, the barometer tube is surrounded throughout by a brass case.

iv) Main scale & vernier scale: Barometer normally consists of two type scale, one millibars and another mm Hg. Main scale of barometer is graduated in whole millibars usually from 870mb to 1100mb at the left of the slot and mm Hg at the right of the slot in the case.

The vernier scale is graduated from 0to 9. 10 vernier divisions exactly cover 39 millibar divisions. It is thus possible to read directly to 0. 1mb.Vernier scale is adjusted by means of a rack and pinion movement.

v) Attached thermometer: Mercury in glass thermometer is mounted in a metal frame on the front of the barometer to indicate the temperature of instrument.

2.7 Rain Measuring Instruments

Precipitation: The total amount of precipitation which reaches the ground in a stated period is expressed as the depth to which it would cover a horizontal projection of the earth's surface if there were no loss of the water by evaporation, runoff.

Requirement of a rain gauge: The most important requirements of a gauge are as follows:

a. The rim of the collector should have a sharp edge and should fall away vertically on the inside and be steeply levelled on the outside.
b. The area of the orifice should be known to the nearest 0.5 per cent, and the construction should be such that this area remains constant while the gauge is in normal use.

c. The collector should be designed to prevent rain from splashing in and out. This can be achieved if the vertical wall is sufficiently deep and the slope of the funnel is sufficiently steep (at least 45 percent).
d. The construction should be such as to minimize wetting errors.
e. The container should have a narrow entrance and be sufficiently protected from radiation to minimize the loss of water by evaporation.

Exposure of rain gauge: To ensure representative observation, the following environmental conditions should be considered as far as possible:

a) The airflow around the rain gauge should be as horizontal as possible. Avoid sites that are concave, elevated or tilted. Choose a site far from precipices or mountain ridges, where local winds are strongly distorted. Avoid sites where wind blows through or stagnates. Building rooftops should not be considered.
b) Choose sites away from other instruments, trees or buildings. Ideally, the instrument should be installed at a distance from such objects equivalent to at least two to four times their height.
c) As the wind speed near the ground increases with height, the efficiency of precipitation collection decreases the higher a gauge is placed. Accordingly, the receptacle should be placed as low as possible. However, too low a setting will result in the entry of splashed rainwater from the ground or the introduction of ground snow in the case of a snowstorm.
d) The ground surface around the rain gauge should be flat and covered with short grass (lawn) or gravel to prevent raindrops from splashing into the unit from outside.

Rain gauge: There are three types rain gauge are used for rainfall measurement. They are-

1) Ordinary rain gauge.
2) Self-recording rain gauge.
3) Tipping bucket rain gauge.

Ordinary rain gauge & Self-recording rain gauge are mainly used in Bangladesh Meteorological Department.

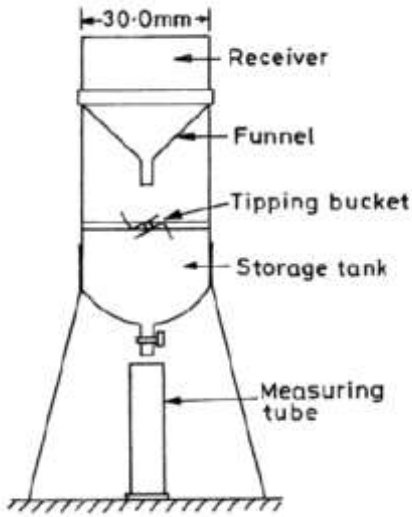
Standard 8-inch ordinary rain gauge: It consists of a 2 feet tall outer cylinder made of a noncorrosive metal, a funnel of the same metal with a brass rim of 8-inch diameter and an

inner tube or container made of brass. The outer cylinder is placed in a metal frame fixed to the concrete base.

Method of measurement:

a) Remove the funnel and insert the measuring stick into the tube taking care that the stick is vertical. To ensure this see that the stick touches one side of the tube both at its top to bottom. Then take the stick out and read the graduation on the stick up to which it has been wetted by water level in the tube.
b) Next take out the tube and threw away its contents. Check the body of the rain gauge to see if any water has overflowed into it; if so, pour out its contents into the tube and measure it. Add the two quantities together to get the total rainfall.

Self-recording rain gauge: In this type of recorder, the recording mechanism is set up inside the container which consists of (a) funnel (b) the base casting and (c) the base. The brass rim at the top of the cover and funnel is 8-inches in diameter and rain is led from there via the funnel to a float chamber containing a metal float. The float has a rod which runs in guides in the top and the bottom of the float chamber, protruding through the top of the float chamber, carries the pen which records on a chart fixed on a revolving drum.



The discharge tube of the siphon is inside and coaxial with the tube connecting with the float chamber. The top of this outer tube is a polished glass cap and the discharge tube comes up to a very short distance of this. When the water in the outer tube rises to the top and flows over bend, capillary action causes all the air to be pushed out and down the delivery tube so that a full flow of water starts at once. Similarly, at the end of the siphoning, once air gets to the top of the tube, the siphoning is stopped immediately. When siphoning occurs, the pen falls vertically to zero line of the chart.

2.7.1 Tipping Bucket Rain Gauges

This type of rain gauge generates an electric signal (i.e., a pulse) for each unit of precipitation collected, and allows automatic or remote observation with a recorder or a counter. The only requirement for the instrument connected to the rain gauge is that it must be able to count pulses. Thus, a wide selection of configurations and applications is possible for this measuring system. Solid precipitation can also be measured if a heater is set at the receptacle.

Structure: This type of rain gauge consists of a receiver and a measuring part, with the receiver serving as the container for the device. The measuring part consists of a tipping bucket and a pulse-generating reed switch (or mercury switch) assembled within the receiver. Mercury switch is connected to these tipping buckets to generate an electrical signal (i.e., a pulse) each time the buckets tip.

①: Orifice
②: Shell
③: Funnel
①-③: Receiver
④: Filter net (large)
⑤: Filter net (small)
⑥: Water filter
⑦: Tipping bucket
⑧: Tipping bucket stopper pad stone
⑨: Tipping bucket stopper screw
⑩: Bearing and cover
⑪: Balancing weight
⑫: Base plate
⑬: Bearing-support plate
⑭: Reed-relay switch
⑮: Drain cylinder
⑯: Splash-guard net
⑰: Mosquito net
⑱: Base stand
⑲: Fixing leg

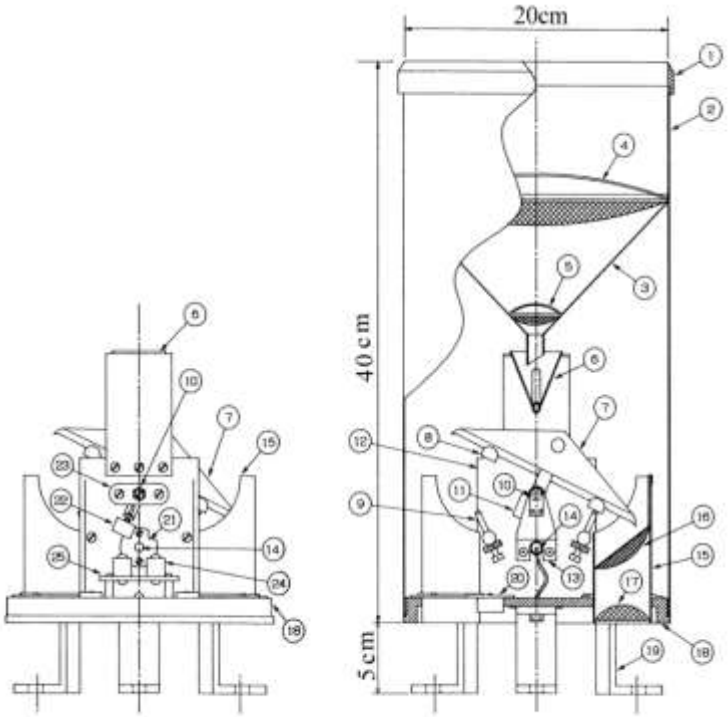


Figure 6.8 Tepping bucket rain gauge

The tipping buckets consist of two triangular vessels attached to the left and right of a rotation shaft, each with a capacity equivalent to a specific amount of precipitation. The reed or mercury switch is connected to these tipping buckets to generate an electrical signal (i.e., a pulse) each time the buckets tip.

Operation: Rainwater collected in the receptacle is channeled through the funnel and poured into a tipping bucket. When it reaches a predetermined amount, the bucket tips and dumps the water into a drain cylinder, causing the reed switch to generate a pulse. Subsequent rainwater is poured into the other bucket. As long as precipitation continues, this operation is repeated, and a pulse is generated each time a bucket tip.

2.8 Temperature Measuring Instruments

Temperature: Temperature is defined to be the thermal condition of any body. The temperature of a body is the condition which determines its ability to communicate heat to the other bodies or received heat from them. In a system of two bodied that which loses heat to the other is said to be at the higher temperature.

Two type temperature measuring instruments are used in BMD. They are-

1. Thermometer.
2. Thermograph.

2.8.1 Type of Thermometer

There are mainly two type of thermometer used in BMD. They are-

1. Liquid- in- glass thermometer.
2. Electric Thermometer (Platinum resistance thermometer).

Type of Liquid in glass thermometer:

1. Mercury- in- glass thermometer.
2. Alcohol- in- glass thermometer.

Liquid- in- glass thermometers are divided into four type with respect their construction, They are-

(a) The sheathed type with the scale engraved on the thermometer stem.

(b) The sheathed type with the scale engraved on an opal glass strip attached to the thermometer tube inside the sheath.

(c) The unsheathed type with the graduation marks on the stem and mounted on a metal, porcelain or wooden back carrying the scale numbers.

(d) The unsheathed type with the scale engraved on the stem.

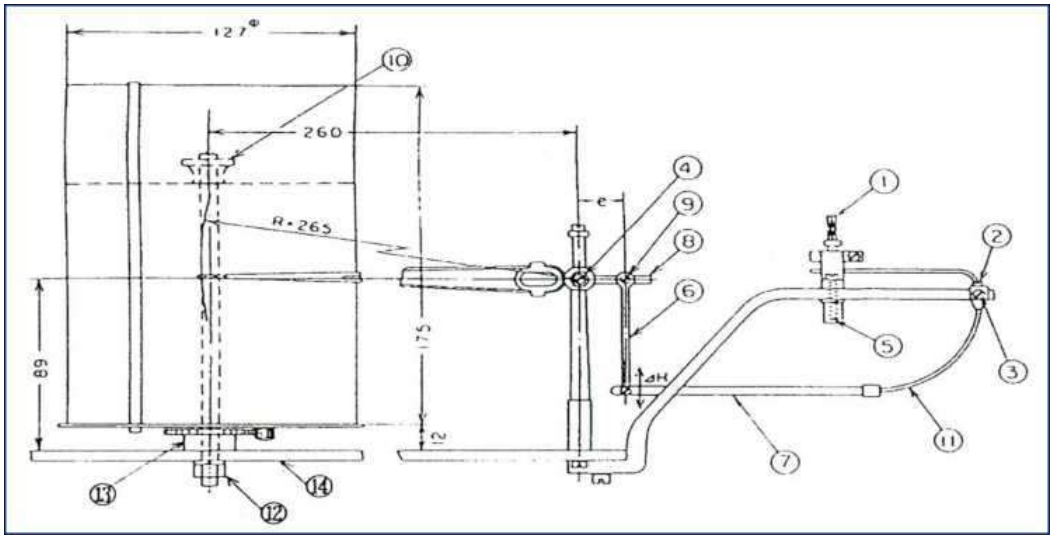
The flowing thermometers are used in BMD:

1. Ordinary thermometer.
2. Maximum thermometer.
3. Minimum thermometer.
4. Grass minimum thermometer.
5. Solar maximum thermometer.
6. Soil thermometer.
7. Floating maximum minimum thermometer.
8. Submersible thermometer.

2.8.2 Thermograph

Main parts of Bimetallic Thermograph:

1) Bi-metal Sensor. (2) Adjusting screw nmmm (3) Pen arm (4) Clock drum (5) lever system & (6) pivot point.



dicator adjustment screw @ Sensor attachment screw ©Stopping screw ® Nut to stop pen arm © Spring © Steel strip © Bimetal lever ® Magnification adjustment ® Screw to stop ® Screw pushing clock-driven drum © Bimetal © Attachment nut for clock-driven drum axis© Washer ® Stand table

Unit: mm

Thermographs are generally used to keep continuous record of air temperature for climatological purposes.

Principle of Bimetallic Thermograph:

The bimetallic element is a strip of metal usually curved or in the form of helix and is made by welding together two bars of metals having different coefficient of expansion. One end of the helix is fixed to the metal frame of the thermograph while the other end is free to move.

As the temperature changes the free end of the helix, to which is connected the pen arm through a mechanism of lever, rotates around its axis and at the same time moves the pen. The instrument should be placed inside a Stevenson's screen.

2.8.3 Using method of Thermograph:

i) Remove the pen from the chart.
ii) Take out the old chart.
iii) Mount a new chart.
iv) Wind the clock, if required.
v) Replace the pen on the chart and leave the instrument for ^ hour or so to reach equilibrium with surrounding.

vi) Then read the dry bulb thermometer and adjust arm until the indicated temperature on the chart is the same as that of the dry bulb thermometer. The adjustment should be done, roughly at first, by loosening the screw holding the pen arm to the spindle, moving the pen arm to approximately correct position and then tightening the screw. Fine adjustment should then be done by use of the setting screw. After making the adjustment once, it should be left for further period undisturbed, and then the result checked. Make further adjustment, if necessary.

2.9 Wind Measuring Instrument

Wind Vane:

Wind direction is estimated with the help of wind vane. It is in effect a balanced lever which turns freely about a vertical axis. In the most common type one end of the lever holds a vain to the wind, whilst the other end is narrow and points to the direction from which the wind is blowing. Under this movable lever is a fixed eight cross arm or direction indicators which are set to 8 cardinal directions of N, NE, E, SE, S, SW, W, NW& N.

Most wind vanes do not, however, respond to changes of wind direction when the wind speed is less than 3 knots. The direction of the wind is to be estimated to the 36 points of compass for reporting in synoptic message.

Procedure of wind direction observation from a wind vane:

a) Before taking observation make sure that the wind vane moves freely; give a turn to the vane and allow it to take up the direction of the wind.
b) The wind should be watched for a period of 10 minutes to get the mean wind direction.
c) Always verify the wind direction given by the vane with that estimated by you by other means.

4 types Anemometer (wind speed measuring instrument) are used in BMD:

1. Cup counter anemometer.
2. Pressure tube anemograph (PT).
3. Mechanical wind recorder.
4. Electrical anemometer.

Exposure of wind instrument:

A suitable site must be free from eddies of trees, buildings hills etc. The standard exposure of the wind instrument is internationally agreed to be over open level terrain at a height of 10

meter. Open terrain is defined as an area where the horizontal distance between the instrument and an obstacle is at least ten times of the height of the obstacle above the ground.

Cup counter Anemometer:

The common type of instrument in use for measuring wind speed is the cup-counter anemometer. It consists of 3 or 4 hemispherical cups attached to the ends of cross metal arms. The cross is pivoted at its central point to a vertical spindle passing through a brass tube attached to the anemometer box. The foot of the spindle rests on a steel ball placed inside ahollow at the base of the box. The rotation of the upright spindle is transferred by means of gear to a counter called cyclometer. From the cyclometer reading wind speed is calculated.

Method of observation and calculation of speed:

1. Record the first cyclometer reading.
2. Take the second reading after 10 minutes and record it in the data card.
3. Calculate the speed as follow:

2nd reading of anemometer ….. 0853.47

1st reading of anemometer ….. 0852.14

Difference = 1.33

Wind speed = $\frac{1.33}{10} \times 60$ Knots

= 7.98 knots

= 8knots

Units

kt	m/s	km/h	mph	ft/s
1.000	0.515	1.853	1.152	1.689
1.943	1.000	3.600	2.237	3.281
0.868	0.447	1.609	1.000	1.467
0.540	0.278	1.000	0.621	0.911
0.592	0.305	1.097	0.682	1.000

A number of different units are used to indicate wind speed, including meters per second (m/s), kilometers per hour (km/h), miles per hour (mph), feet per second (ft/s) and knots (kt). in synoptic reports, the average wind speed measured over a period of 10 minutes is reported in knots (kt). table 4.1 shows the conversion for these units.

Pressure Tube Anemograph (PT anemograph):

Pressure Tube Anemograph consists of (i) Head and Vane (ii) Mastcarrying the direction transmitting rod and pressure suction pipes and supporting the head, (iii) speed and direction recording unit.

Parts of PT Anemograph:

Changing the chart of PT Anemograph:

i. Close the stop cocks and note that the exact time (UTC) to nearest minute. Allow the speed pen to come to rest without touching it, but if the air is calm, raise the float slightly or allow it to fall so that the vertical time mark is produced by the speed pen.

1. Vane.	9. Locking screw.	17. Clock cylinder.
2. Head	10.One- point clutch	18. Clock.
3. Mast.	11. Locking ring.	19. Gauge.
4. Pressure tube.	12. Pen arm.	20. Little stop cork.
5. Suction tube.	13. Socket screw.	21. Pointer.
6. Stop cork	14.Shot	22. Helex
7. Milled head screw.	15. Shot cup.	23. Direction rod.
8. Lower milled head.	16. Float rod.	

ii. Rotate the clock cylinder slightly, so that the speed pen marks its resting position upon the chart.

iii. Swing the three pens away from the chart by means of the pen lifter. Remove the completed chart and insert a blank chart in position taking usual precaution as required for mounting a chart on the clock drum.

iv. Fill the pen with ink after taking care that they are clean. For cleaning, lift them out of their sockets and wash them, but before removing the direction pen arms from their sockets open the one-point clutch to prevent the cam jamming in the helix. Spirit should be used to wash the pen.

v. Swing the two direction pens into contact with the chart, so that they indicate approximately the correct time (UTC).If necessary, raise or lower the clock cylinder by means of the milled headed screw inside the cylinder until upper or lower pen rest exactly upon the upper or lower north line of the chart. The centre screw is the locking screw; the lower milled head is the one which raises or lower the cylinder.

vi. There should be clear evidence upon each chart that both the resting pen and the acting pen will come to rest upon their respective north resting line. This can be secured as follows; open the one-point clutch or raise the locking ring and turn the helix through a complete circle in such a way that the acting pen rises or falls to its resting line. If the acting pen is upper pen arm turn the helix clockwise as viewed from above. If the acting pen is the lower pen arm the helix should be turned in the opposite direction. Close the one-point clutch after the operation direction. Close the one-point clutch after the operation.

vii. Add shot to or remove from the shot-cup until the float just falls to the bottom of the float chamber when both stop clocks are closed and the float is raised about ½ inch. Then, swing the speed pen on the chart. If necessary, adjust the speed pen to fall exactly upon the zero line of the speed record by means of its socket screw, not by raising or lowering the clock cylinder.

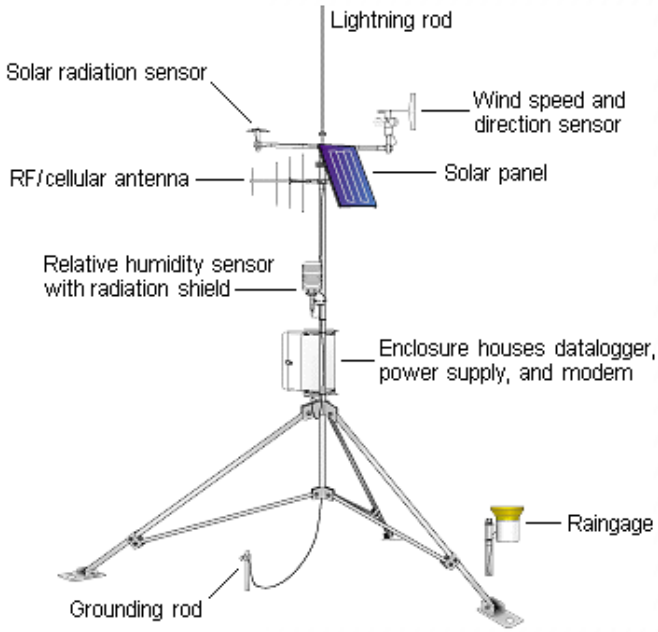
Maintenance of PT Anemograph:

i. Utmost care should be given to general cleanliness of the instrument. Ink should not be allowed to solidify on any part of the pen. Any ink drops that fall on any part of the instrument should be cleaned properly.
ii. To prevent sticking of the speed pen, the float rod must be kept clean and dry. No metal polish, lubricating oil or cleaning material should be used but the rod may be rubbed with a piece of blotting paper previously treated with lead pencil. The gauge should be examined once a fortnight and corrected if necessary. Water may be lost by evaporation and rainwater may find its way down the air tube. Distilled water should be added by removing the milled cap of the gauge tube. Any excess should be run off at the little stopcock just below. The level is correct when the top of the pointer is just in the surface of the water, the stop cocks being closed.
iii. The draining plug by lower cock should be removed from time to time to allow any trapped water to escape. The plug must be screwed firmly back in place.
iv. Take care that the speed pen and the direction pens must be upon the same vertical hour line.
v. The direction mechanism should have a drop of oil about once a month in the hole at the top of the helix, and the bearings of the long levers may need oiling at about the same time. The surfaces of the helixes should be wiped over with an oily rag.

2.10 Automatic Weather Station

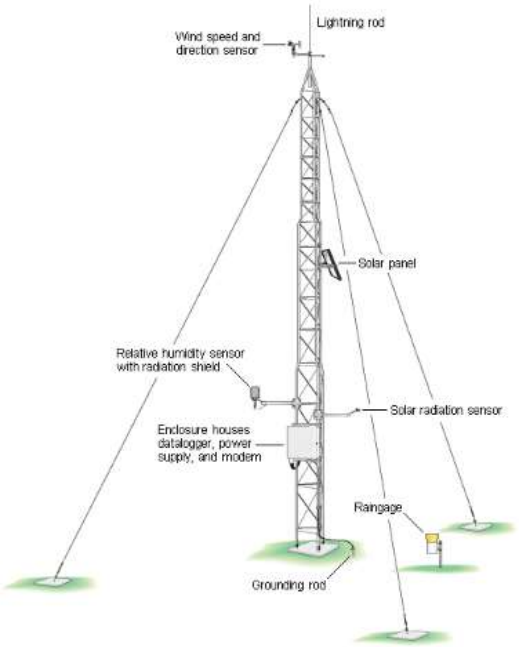
2.10.1 Automatic Weather Stations

Most of the elements required for synoptic, climatological or aeronautical purposes can be measured by automatic instrumentation. As the capabilities of automatic systems increase, the ratio of purely automatic weather stations to observer-staffed weather stations (with or without automatic instrumentation) increases steadily. The guidance in the following paragraphs regarding siting and exposure, changes of instrumentation, and inspection and maintenance apply equally to automatic weather stations and staffed weather stations.



Dataloggers for weather monitoring

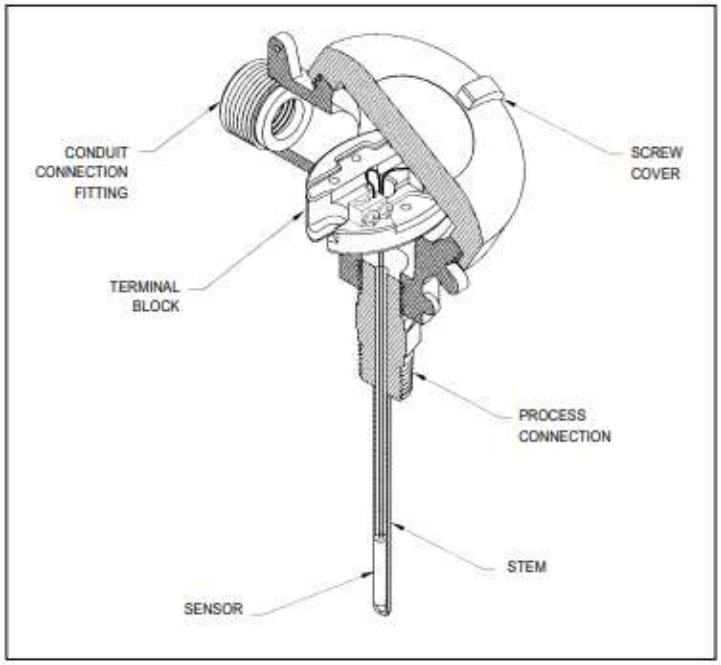
This weather stations are based around a programmable datalogger (typically a CR1000 or CR3000) that measures the sensors, then processes, stores, and transmits the data. Our dataloggers have wide operating temperature ranges, on-board instructions, programmable execution intervals, and ample input channels for commonly used sensors. Wind vector, wet bulb, histogram, and sample on maxima or minima are standard in the datalogger instruction sets. Most sensors can be measured directly—without external signal conditioning.



Data are typically viewed and stored in the units of your choice (e.g., wind speed in mph, m/s, knots). Measurement rates and data recording intervals are independently programmable, allowing calculation of 15-minute, hourly, and daily data values from 1-minute or 1-second measurements, for example. Conditional outputs, such as rainfall intensity and wind gusts, can also be recorded. The program can be modified at any time to accommodate different sensor configurations or new data processing requirements. If needed, channel capacity can be expanded using multiplexers, including a model designed specifically for thermocouples.

2.10.2 Temperature Sensors

A temperature sensor is a device, typically, a thermocouple or RTD, that provides for temperature measurement through an electrical signal. A thermocouple (T/C) is made from two dissimilar metals that generate electrical voltage in direct proportion to changes in temperature. An RTD (Resistance Temperature Detector) is a variable resistor that will change its electrical resistance in direct proportion to changes in temperature in a precise, repeatable and nearly linear manner.



Thermocouples

A thermocouple is made from two dissimilar metal wires. The wires are joined together at one end to form a measuring (hot) junction. The other end, known as the reference (cold) junction, is connected across an electronic measurement device (controller or digital indicator). A thermocouple will generate a measurement signal not in response to actual temperature, but in response to a difference in temperature between the measuring and reference junctions. A small ambient temperature sensor is built into the electronic measuring device near the point where the reference junction is attached. The ambient temperature is then added to the thermocouple differential temperature by the measuring device in order to determine and display the actual measured temperature. Only two wires are necessary to connect a thermocouple to an electrical circuit; however, these connecting wires must be made from the same metals as the thermocouple itself. Adding wire made from other materials (such as common copper wire) will create new measuring junctions that will result in incorrect readings.

RTDs

To greater or lesser degrees, all electrical conducting materials have some amount of resistance to the flow of electricity. When a known electric voltage passes through a conductor, the resistance varies based on the temperature of the conductor. This resistance can be measured and will correspond to a specific temperature. While various elements are affected by temperature in different ways, platinum is commonly used in an RTD due to its purity, linearity and stability over a wide range of temperatures. An electronic readout device, such as a controller or digital indicator designed to measure resistance, is required for use with RTD sensors. Only two standard copper wires are necessary to connect an RTD to an electrical circuit, however, these connecting wires are also subject to small changes in resistance based on surrounding temperature. For this reason, an "extra" third hookup wire is built into most RTDs as a compensation wire to allow the controller or display unit to correct for these variations.

Style

Trerice Temperature Sensors are available in a variety of styles. The weatherproofed screw cover style provides an electrical conduit connection and can be used to house a transmitter (optional). For open system sensing, a nonthreaded style is offered. This design is provided with integrated leadwire and can be Teflon covered to protect the stem and leadwire against corrosive environments. A standard plug with a mating jack may also be furnished.

Stem (Sheath)

All Trerice Thermocouples and RTDs are furnished with a 316 stainless steel stem, with the internal wiring packed in powdered ceramic. The screw head cover style is available in two stem types: welded and spring loaded. The welded stem is suitable for use in liquid applications. The spring-loaded stem is designed to bottom out inside a thermowell, providing

maximum heat sensitivity. Spring loaded stems are not pressure tight and may allow process media to escape; therefore, they must always be installed in a thermowell.

Insertion (U) Length

The insertion (U) length of a thermocouple or RTD represents its depth into the process vessel or thermowell. Trerice Thermocouples and RTDs are available in standard U-lengths from 2" to 24". Other lengths are available upon special order, please consult factory.

Measuring (Hot) Junction

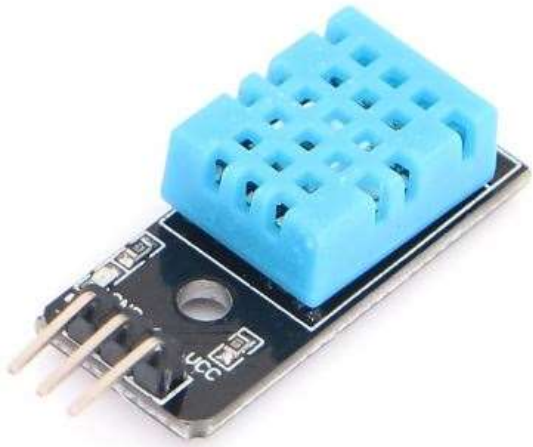
Trerice Thermocouples are available in Type J and Type K and use ceramic insulation to provide an ungrounded measuring junction. Other thermocouple types may be available, please consult factory. Trerice RTDs are a platinum, 3-wire design, and are furnished with either 100Ω or 1000Ω resistance at 32°F (0°C), and a temperature coefficient of 0.00385 Ω/Ω/°C.

Connection (Termination)

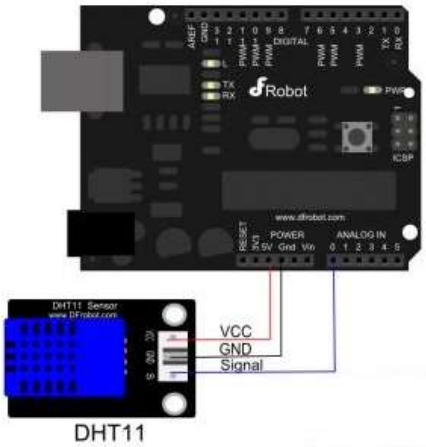
Trerice Thermocouples are provided with terminal block (screw cover head), mating jack, or integrated leadwire connections. The terminal block connection has no leadwire, therefore extension wire must be attached and routed to the electronic measuring device. Thermocouple extension wire must be identical to the thermocouple type, otherwise multiple measuring junctions will be made, causing inaccurate temperature readings. Trerice RTDs are provided with a terminal block (screw cover head) or integrated leadwire connection. The terminal block connection has no leadwire, therefore extension wire must be attached and routed to the indicator or controller.

2.10.3 Temperature and Humidity Sensor

This DHT11 Temperature and Humidity Sensor features a calibrated digital signal output with the temperature and humidity sensor capability. It is integrated with a high-performance 8-bit microcontroller. Its technology ensures the high reliability and excellent long-term stability. This sensor includes a resistive element and a sensor for wet NTC temperature measuring devices. It has excellent quality, fast response, anti-interference ability and high performance.



Each DHT11 sensors features extremely accurate calibration of humidity calibration chamber. The calibration coefficients stored in the OTP program memory, internal sensors detect signals in the process, we should call these calibration coefficients. The single-wire serial interface system is integrated to become quick and easy. Small size, low power, signal transmission distance up to 20 meters, enabling a variety of applications and even the most demanding ones. The product is 4-pin single row pin package. Convenient connection, special packages can be provided according to users need.



DHT11
Hunidity and temperature sensor

Specification

- Supply Voltage: +5 V
- Temperature range :0-50 °C error of ± 2 °C
- Humidity :20-90% RH ± 5% RH error
- Interface: Digital

2.10.4 Present Weather Sensor:

Present Weather Sensor is essentially the FD12 Visibility Meter with the addition of a 'Present Weather Sensor Kit'. The kit comprises of a rain detecting capacitive sensor (circled in the photo), temperature sensor and associated electronics and software. The capacitive sensing device is used to measure the water content from precipitation. This information is coupled with scatter signal data from the optical sensor to estimate the basic precipitation type and intensity. The temperature sensor also provides input to determine the weather code output.


Vaisala FD12P Present Weather Sensor

The FD12P can also detect the presence of non-precipitation phenomena such as fog, mist and haze.

The present weather field in the METAR/SPECI AUTO message at some non-staffed locations throughout the country is being populated by data derived from the FD12P.

SESSION 3: CLOUDS AND PRECIPITATION

The most important condensation processes in the free atmosphere are associated with adiabatic cooling of ascending air and with redistribution of heat and moisture as a result of turbulent transfer. The cloud forms will, therefore, reveal to the analyst the broad aspects of the processes in operation. While condensation processes are capable of producing cloud droplets (or ice particles) of the order of 40 p in diameter, an ordinary raindrop falls within the range from 1 to 6 mm in diameter. Thus, the mass of a common raindrop is about 105 times larger than that of an ordinary cloud droplet, and this brings to the fore the difference between the condensation and precipitation processes. The purpose of this chapter is to describe the various forms of clouds and precipitations and to give a brief account of the condensation processes and the release of precipitation.

As seen from the earth's surface the clouds may be divided into three main categories:
(1) cirrus, or feathery clouds.
(2) stratus, or layer clouds;
(3) cumulus, or heap clouds.
These basic forms may appear in various combinations, such as layers of cumulus, layers of cirrus, etc. Depending upon such combinations and the height above the ground, the clouds may be divided into 10 principal types, as shown in Table 3.1.

Cirrus (Ci) is the highest cloud. It has a typical fibrous (threadlike) structure and a white delicate and silky appearance. Cirrus clouds are sometimes arranged irregularly in the sky as detached clouds without connection with cirrostratus or altostratus. They are then called fair-weather cirrus. If the cirrus clouds are arranged in bands or connected with cirrostratus or altostratus or otherwise systematically arranged, they are usually harbingering of bad weather. In thundery or squally weather, a special kind of cirrus (cirrus densus) is frequently observed which originates from the anvils of cumulonimbus. These clouds are often called false cirrus, because they are denser and usually lower than the ordinary cirrus.

The following descriptions are summarized from the International Atlas of Clouds and States of Sky, Vol. I, General Atlas, Office Nationale Meteorologique, Paris, 1932. An extended revised atlas is in preparation by the World Meteorological Organization

Cirrostratus (Cs) is a thin, whitish sheet of cloud, sometimes like a veil covering the whole sky and merely giving it a milky appearance, at other times showing signs of a fibrous structure like a tangled web. Cirrostratus often produces a halo around the sun or moon. It is often a sign of approaching bad weather.

Detached cirrus.

Cirrocumulus (Cc) consists usually of small, white flakes of clouds without shadow, arranged in a regular pattern. Cirrocumulus develops from cirrostratus. The pattern is due to a single or a double undulation of the cloud sheet

3.1 Basic Types of Clouds

Name	Symbol	Altitude	Height
Cirrus…… Cirrostratus…… Cirrocumulus……	Ci Cs Cc	High	20000-40000 feet
Altostratus…… Altocumulus……	As Ac	Medium	8000-20000 feet
Stratocumulus…… Nimbostratus……	Sc Ns	Low	Surface-20000 feet
Cumulus…… Cumulonimbus……	Cu Cb	Vertical development	

Cirrostratus with halo

Cirrocumulus

Altostratus (As) is a dense sheet of gray or bluish color, often showing a fibrous structure. It often merges gradually into cirrostratus. Increasing altostratus is usually followed by precipitation of a continuous and lasting type.

Altocumulus (Ac) differs from cirrocumulus in consisting of larger globules, often with shadows, whereas cirrocumulus clouds show only indications of shadows or none at all. Altocumulus often develops from dissolving altostratus. An important variety of altocumulus is called altocumulus castellatus. In appearance it resembles ordinary Ac; but in places turreted tops develop that look like miniature cumulus. Altocumulus castellatus usually indicates a change to a chaotic, thundery sky.

Stratocumulus (Sc) is a cloud layer consisting of large, lumpy masses or rolls of dull gray color with brighter interstices. The masses are often arranged in a regular way and resemble altocumulus.
Nimbostratus (Ns) is a dense, shapeless, and ragged layer of low clouds from which-steady precipitation usually falls. It is usually connected with altostratus which is present above the nimbus. Fragments of nimbus that drift under the rain clouds are called fractonimbus or scud.


Altocumulus


Stratocumulus

Cumulus (Cu) is a thick cloud whose upper surface is dome-shaped, often of a cauliflower structure, the base being usually horizontal. Cumulus clouds may be divided into two classes. Flat cumulus clouds without towers or protuberances. are called cumulus humilis or fair-weather cumulus. Towering cumulus clouds with typical cauliflower structure showing internal motion and turbulence are called cumulus congestus. They may develop into cumulonimbus.
Cumulonimbus (Cb) thunderclouds or shower clouds are great masses of cloud rising like mountains, towers, or anvils and having a base that looks like a ragged mass of nimbostratus. The tops are often anvilshaped or surrounded by false cirrus. The above figure shows a cumulonimbus without anvil, and Fig. 22.1.10 shows one with anvil. The cumulonimbus clouds are accompanied by showers, squalls, or thunderstorms and sometimes hail. The line squall cloud is a variety of cumulonimbus that extends like a long line or arch across the sky.

Stratus (St) is a uniform layer of low cloudlike fog, but not lying on the ground. The heights of the various types of cloud vary within wide limits. The high clouds are usually above 20,000 ft (6000 m) and below 40,000 ft (13,000 m). The medium clouds are most frequently between 8000 and 20,000 ft (2500 and 6000 m), and the low clouds below 8000 ft (2500 m).

The tops of cumulus clouds, notably those of cumulonimbus, may reach to great heights, while their bases on the average are at or below 3000 ft (1000 m) above the ground.

3.2 Genetical Classification of Clouds

From a genetical point of view the clouds may be divided into four categories:

1) clouds that form in unstable air masses,
2) clouds that form in stable air masses,
3) clouds that form in connection with quasi-horizontal inversions, and
4) frontal clouds.

Categories (1) and (2) are typical air-mass clouds, or internal clouds, that reveal the stability conditions of the masses to which they belong. The clouds belonging to (3) are most frequently related to group (2), but sometimes they are in certain features related' to (1). The clouds of group (4) are formed because of upglide motion associated with frontal zones. These clouds, too, are influenced by the stability conditions of the air.

3.3. Cloud Development and Precipitation

3.3.1 Atmospheric Stability

We know that most clouds form as air rises, expands, and cools. When we speak of atmospheric stability, we are referring to a condition of equilibrium. For example, rock A resting in the depression in Fig. 2.1 is in stable equilibrium. If the rock is pushed up along either side of the hill and then let go, it will quickly return to its original position. On the other hand, rock B, resting on the top of the hill, is in a state of unstable equilibrium, as a slight push will set it moving away from its original position. Applying these concepts to the atmosphere, we can see that air is in stable equilibrium when, after being lifted or lowered, it tends to return to its original position- it resists upward and downward air motions. Air that is in unstable equilibrium will, when given a little push, move farther away from its original position- it favors vertical air currents. In order to explore the behavior of rising and sinking air, we must first review some concepts we learned in earlier chapters. Recall that a balloon like blob of air is called an air parcel. When an air parcel rises, it moves into a region where the air pressure surrounding it is lower. This situation allows the air molecules inside to push outward on the parcel walls, expanding it. As the air parcel expands, the air inside cools. If the same parcel is brought back to the surface, the increasing pressure around the parcel squeezes (compresses) it back to its original volume, and the air inside warms. Hence, a rising parcel of air expands and cools, while a sinking parcel is compressed and warms. If a parcel of air expands and cools, or compresses and warms, with no interchange of heat with its outside surroundings, this situation is called an adiabatic process. As long as the air in the parcel is unsaturated (the relative humidity is less than 100 percent), the rate of adiabatic cooling or warming remains constant and is about 10°C for every 1000 meters of change in elevation, or about 5.5°F for every 1000 feet. Since this rate of cooling or warming only applies to unsaturated air, it is called the dry adiabatic rate (Fig. 3.2).

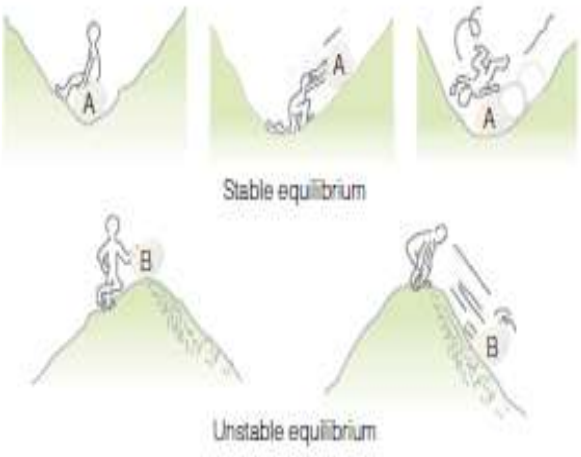


Fig. 3.1: When rock A is disturbed, it will return to its original position; rock B, however, will accelerate away from its original position.

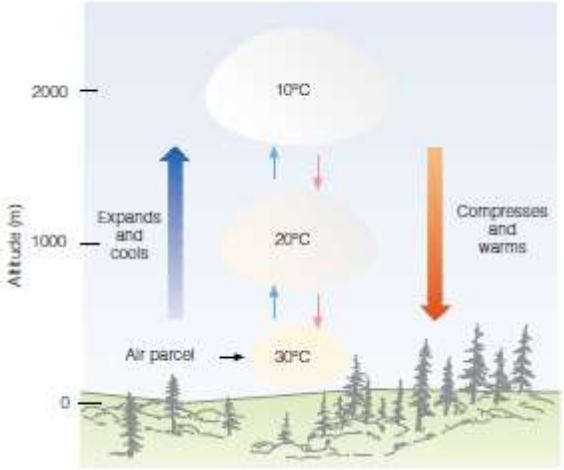


Fig. 3.2: The dry adiabatic rate. As long as the air parcel remains unsaturated, it expands and cools by 10°C per 1000 m; the sinking parcel compresses and warms by 10°C per 1000 m.

As the rising air cools, its relative humidity increases as the air temperature approaches the dew-point temperature. If the air cools to its dew-point temperature, the relative humidity becomes 100 percent. Further lifting results in condensation, a cloud forms, and latent heat is released into the rising air. Because the heat added during condensation offsets some of the cooling due to expansion, the air no longer cools at the dry adiabatic rate but at a lesser rate called the moist adiabatic rate. (Because latent heat is added to the rising saturated air, the process is not really adiabatic). If a saturated parcel containing water droplets were to sink, it would compress and warm at the moist adiabatic rate because evaporation of the liquid droplets would offset the rate of compressional warming. Hence, the rate at which rising or sinking saturated air changes temperature—the moist adiabatic rate—is less than the dry adiabatic rate.

Unlike the dry adiabatic rate, the moist adiabatic rate is not constant, but varies greatly with temperature and, hence, with moisture content—as warm saturated air produces more liquid water than cold saturated air. The added condensation in warm, saturated air liberates more latent heat. Consequently, the moist adiabatic rate is much less than the dry adiabatic rate when the rising air is quite warm; however, the two rates are nearly the same when the rising air is very cold. Although the moist adiabatic rate does vary, to make the numbers easy to deal with we will use an average of 6°C per 1000 m (3.3°F per 1000 ft) in most of our examples and calculations.

3.3.2 Convection and Clouds

Some areas of the earth's surface are better absorbers of sunlight than others and, therefore, heat up more quickly. The air in contact with these 'hot spots' becomes warmer than its surroundings. A hot 'bubble' of air—a thermal—breaks away from the warm surface and rises,

expanding and cooling as it ascends. As the thermal rises, it mixes with the cooler, drier air around it and gradually loses its identity. Its upward movement now slows. Frequently, before it is completely diluted, subsequent rising thermals penetrate it and help the air rise a little higher. If the rising air cools to its saturation point, the moisture will condense, and the thermal becomes visible to us as a cumulus cloud.

Observe in Fig. 3.3 that the air motions are downward on the outside of the cumulus cloud. The downward motions are caused in part by evaporation around the outer edge of the cloud, which cools the air, making it heavy. Another reason for the downward motion is the completion of the convection current started by the thermal. Cool air slowly descends to replace the rising warm air. Therefore, we have rising air in the cloud and sinking air around it. Since subsiding air greatly inhibits the growth of thermals beneath it, small cumulus clouds usually have a great deal of blue sky between them (Fig. 3.4).

As the cumulus clouds grow, they shade the ground from the sun. This, of course, cuts off surface heating and upward convection. Without the continual supply of rising air, the cloud begins to erode as its droplets evaporate. Unlike the sharp outline of a growing cumulus, the cloud now has indistinct edges, with cloud fragments extending from its sides. As the cloud dissipates (or moves along with the wind), surface heating begins again and regenerates another thermal, which becomes a new cumulus. This is why you often see cumulus clouds form, gradually disappear, then reform in the same spot. The stability of the atmosphere plays an important part in determining the vertical growth of cumulus clouds. For example, if a stable layer (such as an inversion) exists near the top of the cumulus cloud, the cloud would have a difficult time rising much higher, and it would remain as a "fair-weather" cumulus cloud. However, if a deep, conditionally unstable layer exists above the cloud, then the cloud may develop vertically into a towering cumulus congestus with a cauliflower like top. When the unstable air is several miles deep, the cumulus congestus may even develop into a cumulonimbus (Fig. 3.5).

Notice in Fig. 3.6 that the distant thunderstorm has a flat anvil-shaped top. The reason for this shape is due to the fact that the cloud has reached the stable part of the atmosphere, and the rising air is unable to puncture very far into this stable layer. Consequently, the top of the cloud spreads laterally as high winds at this altitude (usually above 10,000 m or 33,000 ft) blow the cloud's ice crystals horizontally.

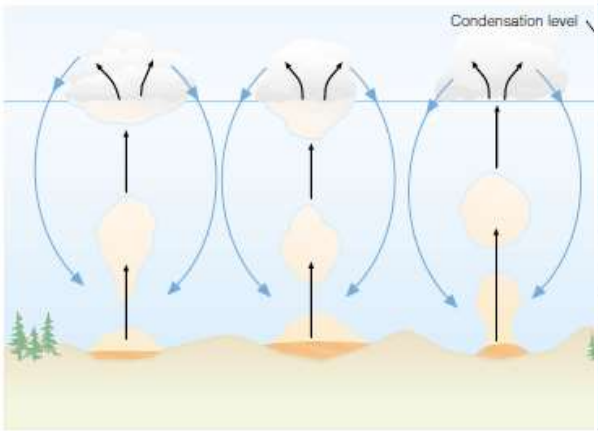


Fig. 3.3: Cumulus clouds form as hot, invisible air bubbles detach themselves from the surface, then rise and cool to the condensation level. Below and within the cumulus clouds, the air is rising. Around the cloud, the air is sinking.

Fig. 3.4: Cumulus clouds developing into thunderstorms in a conditionally unstable atmosphere over the Great Plains. Notice that, in the distance, the cumulonimbus with the anvil top has reached the stable part of the atmosphere.

Fig. 3.5: Cumulus clouds building on a warm summer afternoon. Each cloud represents a region where thermals are rising from the surface. The clear areas between the clouds are regions where the air is sinking.

Fig. 3.6: Cumulus clouds developing into thunderstorms in a conditionally unstable atmosphere over the Great Plains. Notice that, in the distance, the cumulonimbus with the anvil top has reached the stable part of the atmosphere.

3.3.3 Precipitation Processes

As we all know, cloudy weather does not necessarily mean that it will rain or snow. In fact, clouds may form, linger for many days, and never produce precipitation. In Eureka, California, the August daytime sky is overcast more than 50 percent of the time, yet the average precipitation there for August is merely one-tenth of an inch. How, then, do cloud droplets grow large enough to produce rain? And why do some clouds produce rain, but not others? In Fig. 3.7, we can see that an ordinary cloud droplet is extremely small, having an average diameter of 0.02 millimeters (mm), which is less than one-thousandth of an inch. The diameter of a typical cloud droplet is 100 times smaller than a typical raindrop. Clouds, then, are composed of many small droplets- too small to fall as rain. These minute droplets require only

slight upward air currents to keep them suspended. Those droplets that do fall, descend slowly and evaporate in the drier air beneath the cloud.

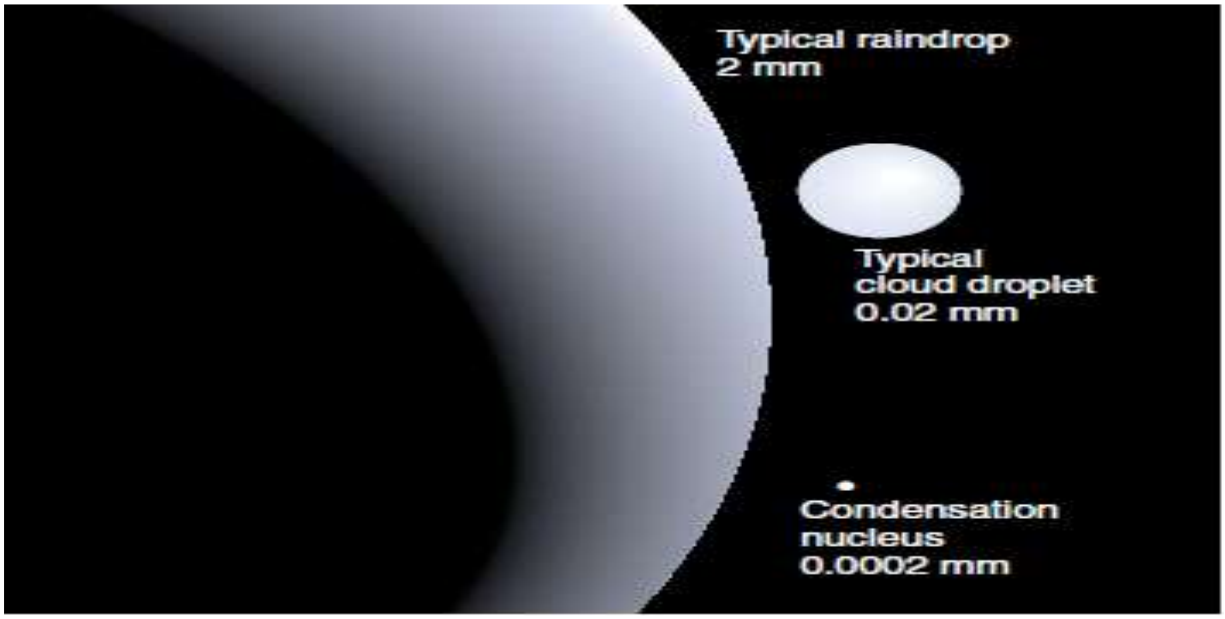


Fig 3.7: Relative sizes of raindrops, cloud droplets, and condensation nuclei.

We know that condensation begins on tiny particles called condensation nuclei. The growth of cloud droplets by condensation is slow and, even under ideal conditions, it would take several days for this process alone to create a raindrop. It is evident, then, that the condensation process by itself is entirely too slow to produce rain. Yet, observations show that clouds can develop and begin to produce rain in less than an hour. Since it takes about 1 million average sizes cloud droplets to make an average size raindrop, there must be some other process by which cloud droplets grow large and heavy enough to fall as precipitation. Even though all the intricacies of how rain is produced are not yet fully understood, two important processes stand out: (i) the collision-coalescence process and (ii) the ice-crystal (or Bergeron) process.

3.3.3.1 Collision and Coalescence Process

In clouds with tops warmer than –15°C (5°F), collisions between droplets can play a significant role in producing precipitation. To produce the many collisions necessary to form a raindrop, some cloud droplets must be larger than others. Larger drops may form on large condensation nuclei, such as salt particles, or through random collisions of droplets. Recent studies also suggest that turbulent mixing between the cloud and its drier environment may play a role in producing larger droplets. As cloud droplets fall, air retards the falling drops. The amount of air resistance depends on the size of the drop and on its rate of fall: The greater its speed, the more air molecules the drop encounters each second. The speed of the falling drop increases until the air resistance equals the pull of gravity. At this point, the drop continues to fall, but at a constant speed, which is called its terminal velocity. Because larger drops have a smaller surface-area-to-weight ratio, they must fall faster before reaching their terminal velocity. Thus, larger drops fall faster than smaller drops. Large droplets overtake and collide with smaller

drops in their path. This merging of cloud droplets by collision is called coalescence. Laboratory studies show that collision does not always guarantee coalescence; sometimes the droplets actually bounce apart during collision. For example, the forces that hold a tiny droplet together (surface tension) are so strong that if the droplet were to collide with another tiny droplet, chances are they would not stick together (coalesce) (Fig. 3.8). Coalescence appears to be enhanced if colliding droplets have opposite (and, hence, attractive) electrical charges. An important factor influencing cloud droplet growth by the collision process is the amount of time the droplet spends in the cloud. Since rising air currents slow the rate at which droplets fall, a thick cloud with strong updrafts will maximize the time cloud droplets spend in a cloud and, hence, the size to which they grow.

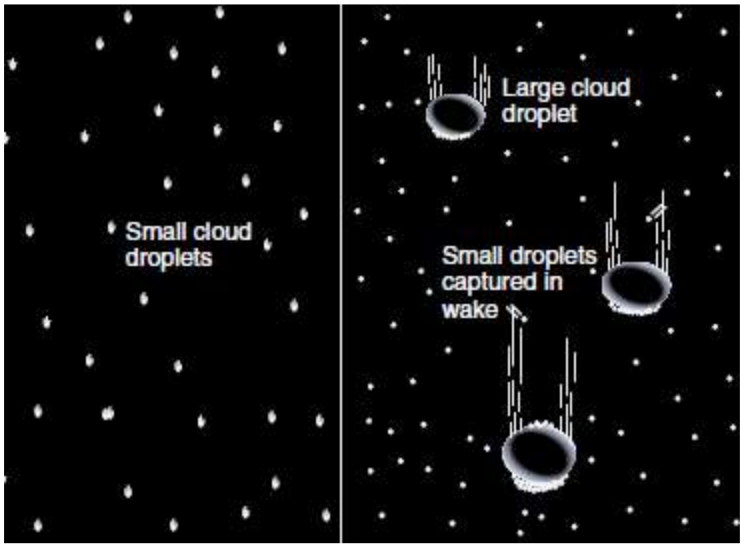


Fig. 3.8: Collision and coalescence. (a) In a warm cloud composed only of small cloud droplets of uniform size, the droplets are less likely to collide as they all fall very slowly at about the same speed. Those droplets that do collide, frequently do not coalesce because of the strong surface tension that holds together each tiny droplet. (b) In a cloud composed of different size droplets, larger droplets fall faster than smaller droplets. Although some tiny droplets are swept aside, some collect on the larger droplet's forward edge, while others (captured in the wake of the larger droplet) coalesce on the droplet's backside.

Clouds that have above-freezing temperatures at all levels are called warm clouds. In tropical regions, where warm cumulus clouds build to great heights, strong convective updrafts frequently occur. In Fig. 3.9, suppose a cloud droplet is caught in a strong updraft. As the droplet rises, it collides with and captures smaller drops in its path, and grows until it reaches a size of about 1 mm. At this point, the updraft in the cloud is just able to balance the pull of gravity on the drop. Here, the drop remains suspended until it grows just a little bigger. Once the fall velocity of the drop is greater than the updraft velocity in the cloud, the drop slowly descends. As the drop falls, some of the smaller droplets get caught in the airstream around it and are swept aside.

Larger cloud droplets are captured by the falling drop, which then grows larger. By the time this drop reaches the bottom of the cloud, it will be a large raindrop with a diameter of over 5 mm. Because raindrops of this size fall faster and reach the ground first, they typically occur at the beginning of a rain shower originating in these warm, convective cumulus clouds.

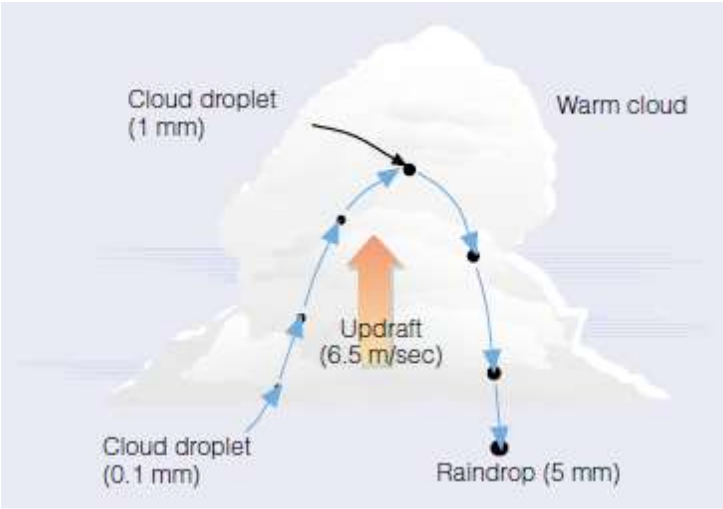


Fig. 3.9: A cloud droplet rising then falling through a warm cumulus cloud can grow by collision and coalescence and emerge from the cloud as a large raindrop.

So far, we have examined the way cloud droplets in warm clouds (that is, those clouds with temperatures above freezing) grow large enough by the collision coalescence process to fall as raindrops. The most important factor in the production of raindrops is the cloud's liquid water content. In a cloud with sufficient water, other significant factors are:

i. the range of droplet sizes;
ii. the cloud thickness;
iii. the updrafts of the cloud;
iv. the electric charge of the droplets and the electric field in the cloud Relatively thin stratus clouds with slow, upward air currents are, at best, only able to produce drizzle (the lightest form of rain), whereas the towering cumulus clouds associated with rapidly rising air can cause heavy showers. Now, let's turn our attention to the ice-crystal process of rain formation.

Relatively thin stratus clouds with slow, upward air currents are, at best, only able to produce drizzle (the lightest form of rain), whereas the towering cumulus clouds associated with rapidly rising air can cause heavy showers. Now, let's turn our attention to the ice-crystal process of rain formation.

3.3.3.2 Ice-Crystal Process

The ice-crystal (or Bergeron*) process of rain formation proposes that both ice crystals and liquid cloud droplets must co-exist in clouds at temperatures below freezing. Consequently, this process of rain formation is extremely important in middle and high latitudes, where clouds are able to extend upwards into regions where air temperatures are below freezing. Such clouds are called cold clouds. Fig. 3.10 illustrates a typical cumulonimbus cloud.

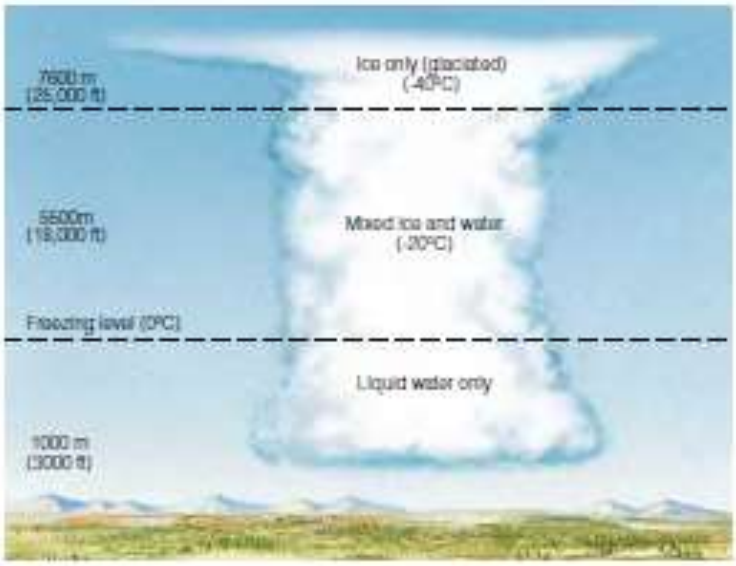


Fig. 3.10: The distribution of ice and water in a cumulonimbus cloud.

In the warm region of the cloud (below the freezing level) where only water droplets exist, we might expect to observe cloud droplets growing larger by the collision and coalescence process described in the previous section. Surprisingly, in the cold air just above the freezing level, almost all of the cloud droplets are still composed of liquid water. Water droplets existing at temperatures below freezing are referred to as supercooled. At higher levels, ice crystals become more numerous, but are still outnumbered by water droplets. Ice crystals exist overwhelmingly in the upper part of the cloud, where air temperatures drop to well below freezing.

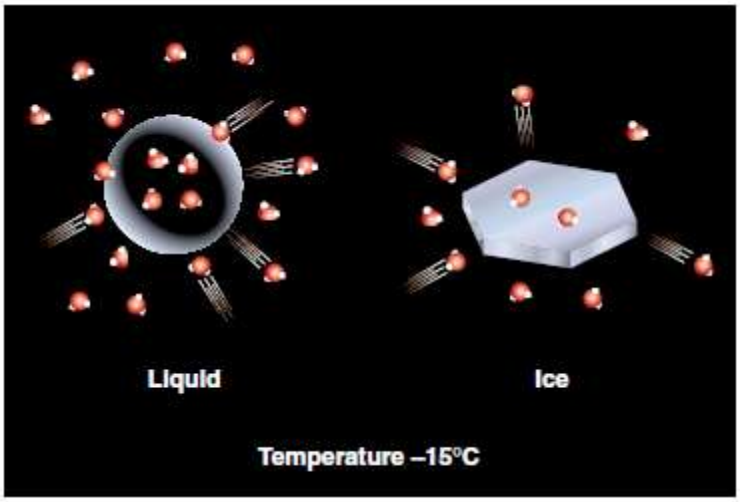


Fig. 3.11: In a saturated environment, the water droplet and the ice crystal are in equilibrium, as the number of molecules leaving the surface of each droplet and ice crystal equals the number returning. The greater number of vapor molecules above the liquid indicates, however, that the saturation vapor pressure over water is greater than it is over ice.

Why are there so few ice crystals in the middle of the cloud, even though temperatures there, too, are below freezing? Laboratory studies reveal that the smaller the amount of pure water,

the lower the temperature at which water freezes. Since cloud droplets are extremely small, it takes very low temperatures to turn them into ice. Just as liquid cloud droplets form on condensation nuclei, ice crystals may form in subfreezing air if there are ice-forming particles present called ice nuclei. The number of ice-forming nuclei available in the atmosphere is small, especially at temperatures above −10°C (14°F). Although some uncertainty exists regarding the principal source of ice nuclei, it is known that certain clay minerals, bacteria in decaying plant leaf material, and ice crystals themselves are excellent ice nuclei.

Moreover, particles serve as excellent ice-forming nuclei if their geometry resembles that of an ice crystal. We can now understand why there are so few ice crystals in the subfreezing region of some clouds. Liquid cloud droplets may freeze, but only at very low temperatures. Ice nuclei may initiate the growth of ice crystals, but they do not abound in nature. Therefore, we are left with a cold cloud that contains many more liquid droplets than ice particles, even at low temperatures. Neither the tiny liquid nor solid particles are large enough to fall as precipitation. How, then, does the ice crystal process produce rain and snow?

In the subfreezing air of a cloud, many supercooled liquid droplets will surround each ice crystal. Suppose that the ice crystal and liquid droplet in (Fig.3.11) are part of a cold (−15°C), supercooled, saturated cloud. Since the air is saturated, both the liquid droplet and the ice crystal are in equilibrium, meaning that the number of molecules leaving the surface of both the droplet and the ice crystal must equal the number of molecules returning.

Observe, however, that there are more vapor molecules above the liquid. The reason for this fact is that molecules escape the surface of water much easier than they escape the surface of ice. Consequently, more molecules escape the water surface at a given temperature, requiring more in the vapor phase to maintain saturation. Therefore, it takes more vapor molecules to saturate the air directly above the water droplet than it does to saturate the air directly above the ice crystal. Put another way, at the same subfreezing temperature, the saturation vapor pressure just above the water surface is greater than the saturation vapor pressure above the ice surface.

This difference in vapor pressure causes water vapor molecules to move (diffuse) from the droplet toward the ice crystal. The removal of vapor molecules reduces the vapor pressure above the droplet. Since the droplet is now out of equilibrium with its surroundings, it evaporates to replenish the diminished supply of water vapor above it. This process provides a continuous source of moisture for the ice crystal, which absorbs the water vapor and grows rapidly (Fig. 3.12).

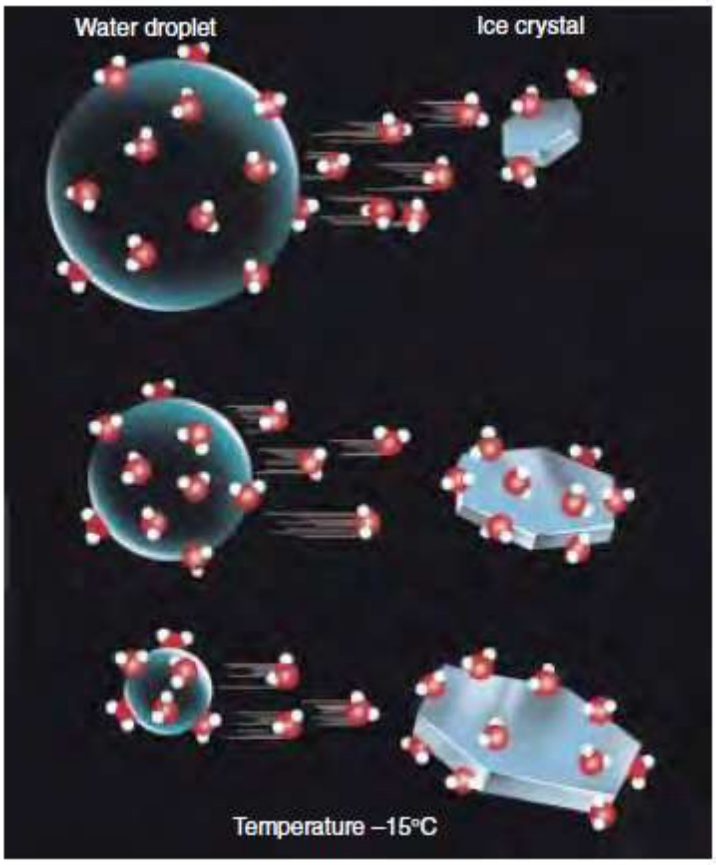
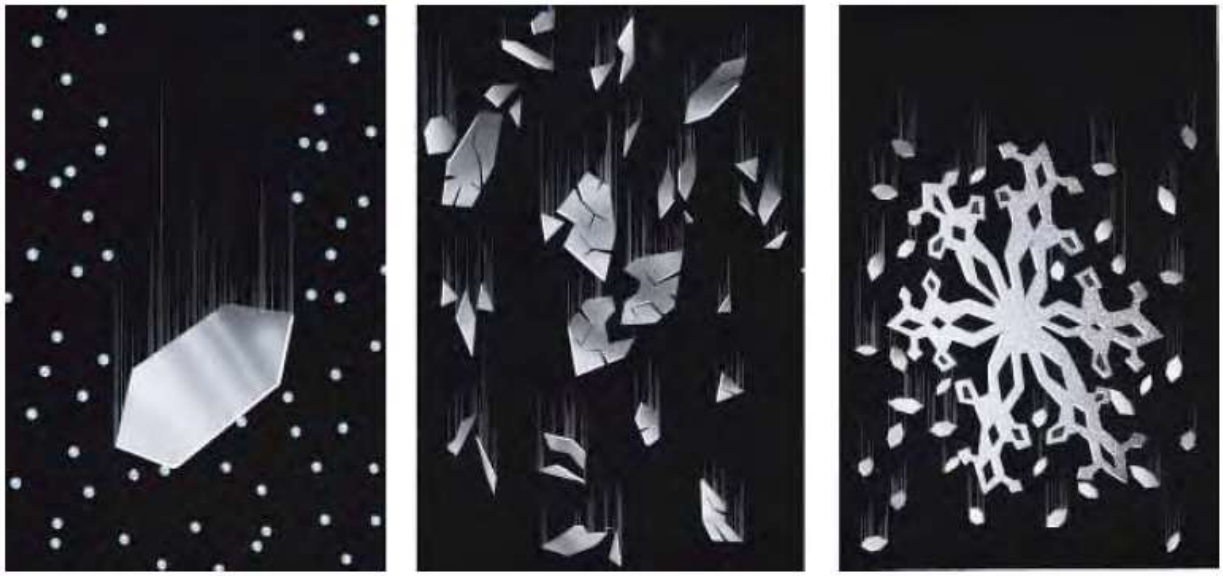


Fig. 3.12: The ice-crystal process. The greater number of water vapor molecules around the liquid droplets causes water molecules to diffuse from the liquid drops toward the ice crystals. The ice crystals absorb the water vapor and grow larger, while the water droplets grow smaller.

Hence, during the ice-crystal (Bergeron) process, ice crystals grow larger at the expense of the surrounding water droplets. The ice crystals may now grow even larger. For example, in some clouds, ice crystals might collide with supercooled liquid droplets. Upon contact, the liquid droplets freeze into ice and stick to the ice crystal—a process called accretion, or riming. The icy matter (rime) that forms are called graupel (or snow pellets). As the graupel falls, it may fracture or splinter into tiny ice particles when it collides with cloud droplets. These splinters may then go on themselves to become new graupel, which, in turn, may produce more splinters. In colder clouds, the delicate ice crystals may collide with other crystals and fracture into smaller ice particles, or tiny seeds, which freeze hundreds of supercooled droplets on contact. In both cases a chain reaction may develop, producing many ice crystals (Fig. 3.13).



(a) Falling ice crystals may freeze supercooled droplets on contact (accretion), producing larger ice particles.

(b) Falling ice particles may collide and fracture into many tiny (secondary) ice particles.

(c) Falling ice crystals may collide and stick to other ice crystals (aggregation), producing snowflakes.

Fig. 3.13: Ice particles in clouds.

As they fall, they may collide and stick to one another, forming an aggregate of ice crystals called a snowflake. If the snowflake melts before reaching the ground, it continues its fall as a raindrop. Therefore, much of the rain falling in middle and northern latitudes—even in summer—begins as snow.

3.3.4 Precipitation Types

Rain: Most people consider rain to be any falling drop of liquid water. To the meteorologist, however, that falling drop must have a diameter equal to, or greater than, 0.5 mm (0.02 in.) to be considered rain.

Drizzle: Fine uniform drops of water whose diameters are smaller than 0.5 mm are called drizzle. Most drizzle falls from stratus clouds; however, small raindrops may fall through air that is unsaturated, partially evaporate, and reach the ground as drizzle.

Virga: Occasionally, the rain falling from a cloud never reaches the surface because the low humidity causes rapid evaporation. As the drops become smaller, their rate of fall decreases, and they appear to hang in the air as a rain streamer. These evaporating streaks of precipitation are called virga (Fig. 3.14).

Cloud burst: Raindrops may also fall from a cloud and not reach the ground if they encounter the rapidly rising air of an updraft. If the updraft weakens or changes direction and becomes a downdraft, the suspended drops will fall to the ground as a sudden rain shower. The showers falling from cumuliform clouds are usually brief and sporadic, as the cloud moves overhead and then drifts on by. If the shower is excessively heavy, it is termed a cloudburst. Beneath a cumulonimbus cloud, which normally contains large convection currents, it is

entirely possible that one side of a street may be dry (updraft side), while a heavy shower is occurring across the street (downdraft side).

Continuous rain, on the other hand, usually falls from a layered cloud that covers a large area and has smaller vertical air currents. These are the conditions normally associated with nimbostratus clouds. Raindrops that reach the earth's surface are seldom larger than about 6 mm (0.2 in.), the reason being that the collisions (whether glancing or head-on) between raindrops tend to break them up into many smaller drops. Additionally, when raindrops grow too large, they become unstable and break apart.

Fig. 3.14: The streaks of falling precipitation that evaporate before reaching the ground are called *virga*.

Fig. 3.15: The dangling white streamers of ice crystals beneath these cirrus clouds are known as fall streaks. The bending of the streaks is due to the changing wind speed with height.

Acid rain: After a rainstorm, visibility usually improves primarily because precipitation removes (scavenges) many of the suspended particles. When rain combines with gaseous pollutants, such as oxides of sulfur and nitrogen, it becomes acidic. Acid rain, which has an adverse effect on plants and water resources, is becoming a major problem in many industrialized regions of the world.

Snow: We have learned that much of the precipitation reaching the ground actually begins as snow. In summer, the freezing level is usually high and the snowflakes falling from a cloud melt before reaching the surface. In winter, however, the freezing level is much lower, and falling snowflakes have a better chance of survival. In fact, snowflakes can generally fall about 300 m (or 1000 ft) below the freezing level before completely melting. When the warmer air beneath the cloud is relatively dry, the snowflakes partially melt. As the liquid water evaporates, it chills the snowflake, which retards its rate of melting. Consequently, in air that is relatively dry, snowflakes may reach the ground even when the air temperature is considerably above freezing.

Although many believe this expression, the fact remains that it is never too cold to snow. True, more water vapor will condense from warm saturated air than from cold saturated air. But, no matter how cold the air becomes, it always contains some water vapor that could produce snow. In fact, tiny ice crystals have been observed falling at temperatures as low as –47°C (–53°F). We usually associate extremely cold air with "no snow" because the coldest winter weather occurs on clear, calm nights—conditions that normally prevail with strong high-pressure areas that have few if any clouds.

When ice crystals and snowflakes fall from high cirrus clouds, they are called fall streaks. Fall streaks behave in much the same way as virga—as the ice particles fall into drier air, they usually disappear as they change from ice into vapor (called sublimation). Because the wind at higher levels moves the cloud and ice particles horizontally more quickly than do the slower winds at lower levels, fall streaks often appear as dangling white streamers (Fig. 3.14). Moreover, fall streaks descending into lower, supercooled clouds may actually seed them.

Snowflakes falling through moist air that is slightly above freezing slowly melt as they descend. A thin film of water forms on the edge of the flakes, which acts like glue when other snowflakes come in contact with it. In this way, several flakes join to produce giant snowflakes that often measure an inch or more in diameter. These large, soggy snowflakes are associated with moist air and temperatures near freezing. However, when snowflakes fall through extremely cold air with a low moisture content, they do not readily stick together and small, powdery flakes of "dry" snow accumulate on the ground. If you catch falling snowflakes on a dark object and examine them closely, you will see that the most common snowflake form is a fernlike branching shape called dendrite (Fig. 3.15).

As ice crystals fall through a cloud, they are constantly exposed to changing temperatures and moisture conditions. Since many ice crystals can join together (aggregate) to form a much larger snowflake, ice crystals may assume many complex patterns. Snow falling from developing cumulus clouds is often in the form of flurries. These are usually light showers that fall intermittently for short durations and produce only light accumulations. A more intense snow shower is called a snow squall. These brief but heavy falls of snow are comparable to

summer rain showers and, like snow flurries, usually fall from cumuliform clouds. A more continuous snowfall (sometimes steadily, for several hours) accompanies nimbostratus and altostratus clouds.

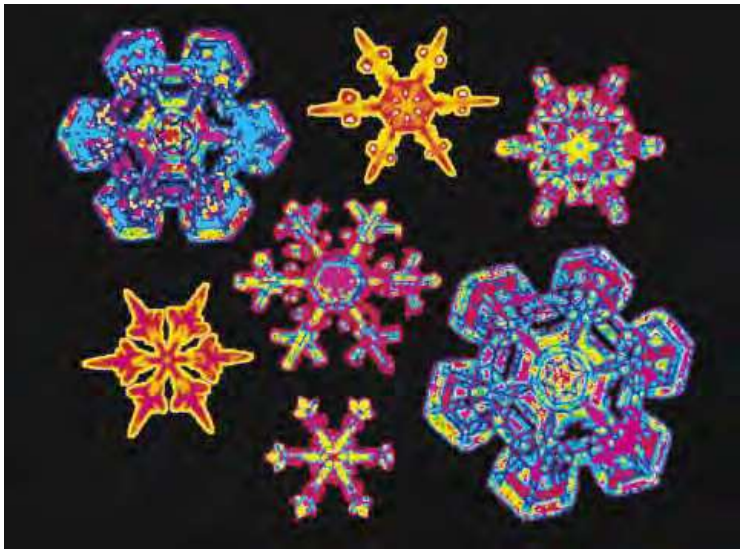


Fig. 3.16: Computer color-enhanced image of dendrite snowflakes.

When a strong wind is blowing at the surface, snow can be picked up and deposited into huge drifts. Drifting snow is usually accompanied by blowing snow; that is, snow lifted from the surface by the wind and blown about in such quantities that horizontal visibility is greatly restricted. The combination of drifting and blowing snow, after falling snow has ended, is called a ground blizzard. A true blizzard is a weather condition characterized by low temperatures and strong winds (greater than 30 knots) bearing large amounts of fine, dry, powdery particles of snow, which can reduce visibility to only a few meters.

Sleet and Freezing rain

Consider the falling snowflake, as it falls into warmer air, it begins to melt. When it falls through the deep subfreezing surface layer of air, the partially melted snowflake or cold raindrop turns back into ice, not as a snowflake, but as a tiny transparent (or translucent) ice pellet called sleet. Generally, these ice pellets bounce when striking the ground and produce a tapping sound when they hit a window or piece of metal. The cold surface layer beneath a cloud may be too shallow to freeze raindrops as they fall. In this case, they reach the surface as supercooled liquid drops. Upon striking a cold object, the drops spread out and almost immediately freeze, forming a thin veneer of ice. This form of precipitation is called freezing rain, or glaze.

If the drops are quite small, the precipitation is called freezing drizzle. When small, supercooled cloud or fog droplets strike an object whose temperature is below freezing, the tiny droplets freeze, forming an accumulation of white or milky granular ice called rime (Fig. 3.16).

Freezing rain can create a beautiful winter wonderland by coating everything with silvery, glistening ice. At the same time, highways turn into skating rinks for automobiles, and the destructive weight of the ice—which can be many tons on a single tree—breaks tree branches, power lines, and telephone cables.

Fig. 3.17: An accumulation of rime forms on tree branches as supercooled fog droplets freeze on contact in the below-freezing air

Fig. 3.18: The Coffeyville hailstone. This giant hailstone—the largest ever reported in the United States—fell on the community of Coffeyville, Kansas, on September 3, 1970. The layered structure of the hailstone reveals that it traveled through a cloud of varying water content and temperature.

3.4 Hail

Hailstones are pieces of ice either transparent or partially opaque, ranging in size from that of small peas to that of golf balls or larger. Some are round, others take on irregular shapes. The largest authenticated hailstone in the United States fell on Coffeyville, Kansas, in September 1970 (Fig. 3.17). This giant weighed 757 grams (1.67 lb) and had a measured diameter of over 14 cm (5.5 in.). Canada's record hailstone fell on Cedoux, Saskatchewan, during August 1973. It weighed 290 grams (0.6 lb) and measured about 10 cm (4 in.) in diameter.

Needless to say, large hailstones are quite destructive as they can break windows, dent cars, batter roofs of homes, and cause extensive damage to livestock and crops. In fact, a single hailstorm can destroy a farmer's crop in a matter of minutes. Estimates are that, in the United States alone, hail damage amounts to hundreds of millions of dollars annually. Although

hailstones are potentially lethal, only two fatalities due to falling hail have been documented in the United States during the twentieth century.

Hail is produced in a cumulonimbus cloud when graupel, large frozen raindrops, or just about any particles (even insects) act as embryos that grow by accumulating supercooled liquid droplets—accretion. For a hailstone to grow to a golf ball size, it must remain in the cloud for between 5 and 10 minutes. Violent, up surging air currents within the cloud carry small embryos high above the freezing level. When the updrafts are tilted, the embryos are swept laterally through the cloud. Studies reveal that the width and tilt of the main updraft are very important to hailstone growth, with the best trajectory being one that is nearly horizontal though the cloud.

As the embryos pass through regions of varying liquid water content, a coating of ice forms around them and they grow larger and larger. When the ice particles are of appreciable size, they become too large and heavy to be supported by the rising air, and they then begin to fall as hail. As they slowly descend, the hailstones may get caught in a violent updraft, only to be carried upward once again to repeat the cycle. Or, they may fall through the cloud and begin to melt in the warmer air below. Small hailstones often melt before reaching the ground, but, in the violent thunderstorms of summer, hailstones may grow large enough to reach the ground before completely melting. Strangely, then, the largest form of frozen precipitation occurs during the warmest time of the year.

As the cumulonimbus cloud moves along, it may deposit its hail in a long, narrow band known as a hail streak. If the cloud should remain almost stationary for a period of time, substantial accumulation of hail is possible. For example, in June 1984, a devastating hailstorm lasting over an hour dumped knee-deep hail on the suburbs of Denver, Colorado. In addition to its destructive effect, accumulation of hail on a roadway is a hazard to traffic as when, for example, four people lost their lives near Soda Springs, California, in a 15-vehicle pileup on a hail-covered freeway in September 1989.

Because hailstones are so damaging, various methods have been tried to prevent them from forming in thunderstorms. One method employs the seeding of clouds with large quantities of silver iodide. These nuclei freeze supercooled water droplets and convert them into ice crystals. The ice crystals grow larger as they come in contact with additional supercooled cloud droplets. In time, the ice crystals grow large enough to be called graupel, which then becomes a hailstone embryo. Large numbers of embryos are produced by seeding in hopes that competition for the remaining supercooled droplets may be so great that none of the embryos would be able to grow into large and destructive hailstones. Russian scientists claim great success in suppressing hail using ice nuclei, such as silver iodide and lead iodide. In the United States, the results of most hail-suppression experiments are still inconclusive.

3.5. Doppler Radar and Precipitation

Radar (radio detection and ranging) has become an essential tool of the atmospheric scientist, for it gathers information about storms and precipitation in previously inaccessible regions. Atmospheric scientists use radar to examine the inside of a cloud much like physicians

use X-rays to examine the inside of a human body. Essentially, the radar unit consists of a transmitter that sends out short, powerful microwave pulses. When this energy encounters a foreign object—called a target—a fraction of the energy is scattered back toward the transmitter and is detected by a receiver. The returning signal is amplified and displayed on a screen, producing an image or "echo" from the target. The elapsed time between transmission and reception indicates the target's distance.

The brightness of the echo is directly related to the amount (intensity) of rain falling in the cloud. So, the radar screen shows not only where precipitation is occurring, but also how intense it is. Typically, the radar image is displayed using various colors to denote the intensity of precipitation within the range of the radar unit.

During the 1990s, Doppler radar replaced the conventional radar units that were put into service shortly after World War II. Doppler radar is like conventional radar in that it can detect areas of precipitation and measure rainfall intensity. Using special computer programs called algorithms, the rainfall intensity, over a given area for a given time, can be computed and displayed as an estimate of total rainfall over that particular area. But the Doppler radar can do more than conventional radar. Because the Doppler radar uses the principle called Doppler shift, it has the capacity to measure the speed at which falling rain is moving horizontally toward or away from the radar antenna. Falling rain moves with the wind. Consequently, Doppler radar allows scientists to peer into a tornado-generating thunderstorm and observe its wind. In some instances, radar displays indicate precipitation where there is none reaching the surface. This situation happens because the radar beam travels in a straight line and the earth curves away from it. Hence, the return echo is not necessarily that of precipitation reaching the ground but is that of raindrops in the cloud.

The Doppler shift (or effect) is the change in the frequency of waves that occurs when the emitter or the observer is moving toward or away from the other. As an example, suppose a high-speed train is approaching you. The higher pitched (higher frequency) whistle you hear as the train approaches will shift to a lower pitch (lower frequency) after the train passes.

SESSION 4: CYCLONES, ANTICYCLONES, AND LOCAL WINDS

4.1 Cyclones and Cyclonic Winds

Although a more complete analysis of the structure of a cyclone will be given in Chap. 12, we may consider for the present that a cyclone is a roughly circular, low-pressure area whose diameter may be from hundreds to a thousand or so miles. Atmospheric pressure is always lowest in the center of this region and increases radially outward.

In Fig. 4-1, the pressure gradient is shown directed inward by the broken arrows. However, the winds in the low are under the influence of the deflective (Coriolis), centrifugal, and frictional forces as well as the pressure gradient. Instead of blowing directly inward, parallel to the gradient, the wind blows across the isobars, at a high angle to the pressure gradient.

In the Northern Hemisphere the deflective force causes the wind to blow to the right of the pressure gradient, resulting in a counterclockwise spiral, inward motion for the air near the surface. At an elevation of thousands of feet above the ground, frictional forces are very low, so that the wind deflection is greater, giving motion parallel to the isobars. Such winds, that move parallel to the isobars as the result of balance between the pressure gradient and the deflective forces, were defined earlier as gradient when motion is curved and geostrophic when it is more or less linear.

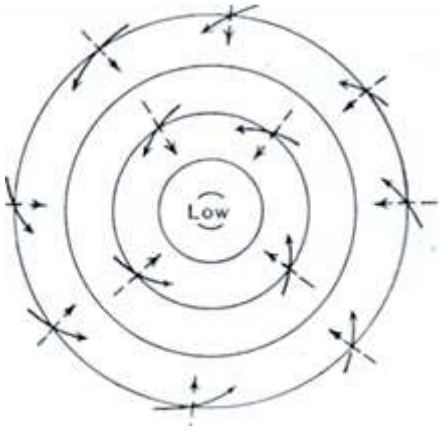


Fig-4.1: Cyclonic circulation in the Northern Hemisphere

The counterclockwise motion in the Northern Hemisphere is known as cyclonic circulation. Note that the term cyclone thus implies not particularly dangerous or destructive storm but is one of the typical and common weather patterns of the middle latitudes. Although steep pressure gradients, with consequent gale winds, often develop in middle-latitude low-pressure areas, the latter should not be confused with the more violent but much smaller tropical cyclones or hurricanes. In distinction to these, the mid-latitude lows are known technically as extratropical cyclones.

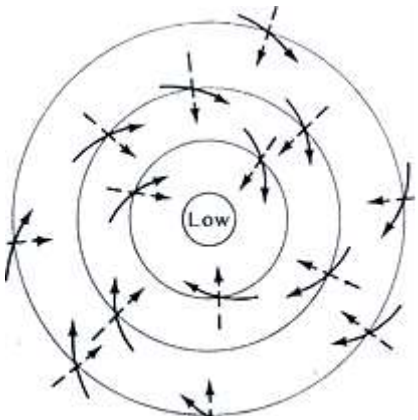
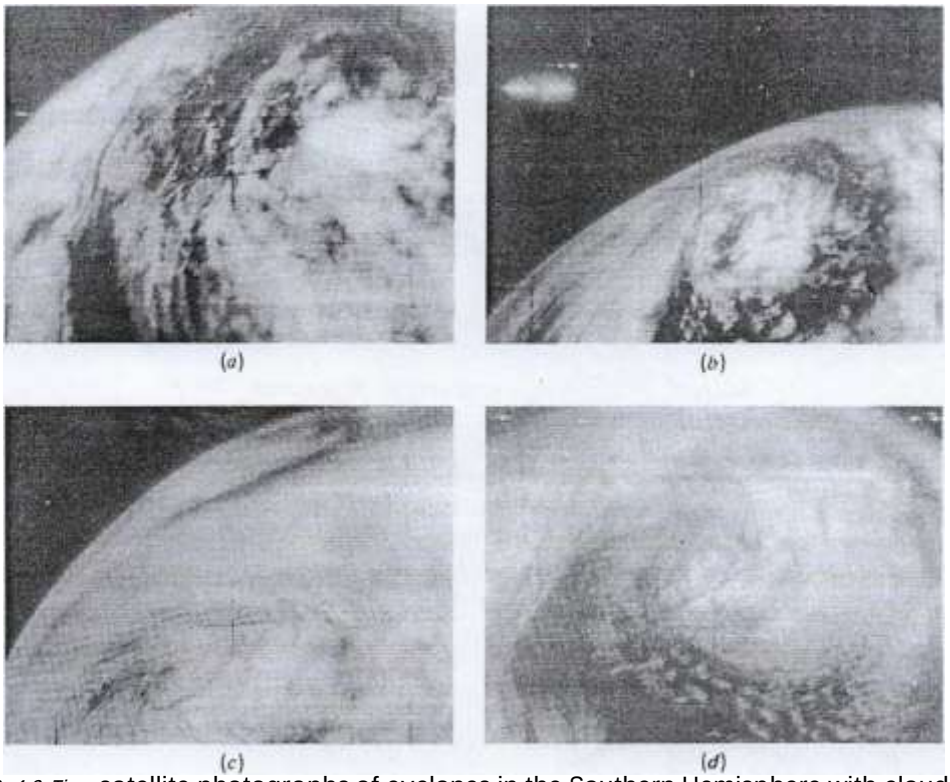


Fig-4.2: Cyclonic circulation in the Southern Hemisphere

Recall that the deflective force is to the left of the pressure gradient in the Southern Hemisphere. Consequently, the wind motion takes the pattern shown in Fig. 4-2, becoming clockwise in circulation about a cyclone in the southern latitudes. Figures 4-3 and 4-5 are striking satellite photographs of cloud distribution in the Northern and Southern Hemispheres, respectively. The former, a Tiros VI photograph, clearly shows the counterclockwise, inward, wind- spiral motion by means of the cloud streaks in the upper right quadrant. The latter indicates clockwise motion for the Southern Hemisphere cyclonic disturbance and shows progressive changes in this storm during an eight-day period. It is instructive to compare the cloud pattern of the cyclone in Fig. 4-3 with the abridged marine weather map of the time, as shown in Fig. 4-4, which shows the isobaric pattern of lows and highs analyzed on the basis of ships' reports. The prominent cloud in the lower half of the photograph is associated with the frontal system shown by the heavy line with black semicircular and triangular symbols.

4.2 Anticyclones

The roughly circular high-pressure areas described in Chap. 8 give rise to winds whose motion with respect to the center is essentially opposite to that developed in low-pressure areas. Under the influence of both the pressure gradient and the deflective forces, the winds, which tend to move directly out from the center of a high-pressure system, develop a clockwise outward spiral in the Northern Hemisphere and a counterclockwise motion in the Southern Hemisphere. These motions are illustrated for each hemisphere, respectively,



4.6. Tiros satellite photographs of cyclones in the Southern Hemisphere with cloud patterns clearly showing the clockwise circulation, *(a)* Tropical cyclone in the Indian Ocean, April 20,1960. *(b)* Indian Ocean on April 25,1960. *(c)* Indian Ocean, April 30,1960. *(d)* South Atlantic Ocean, April 28, 1960. (National Weather Service)

in Figs. 4-7 and 4-8, in which the former indicates the direction of the pressure gradient by dashed arrows. Because the wind circulation in high-pressure systems is opposite that in cyclones (low-pressure systems), the motion is called anticydonic and the system is called an anticydone.

As in the case of cyclones, frictional forces decrease vertically upward. Consequently, the higher-speed winds at levels several thousand feet above the ground suffer the maximum deflection possible, which is a direction normal to the pressure gradient. Hence the wind well above the ground is influenced by the geostrophic or gradient conditions and tends to blow parallel to the isobars, rather than across them at a small angle as near the surface.

Although we usually associate storms and strong winds with cyclonic activity, very strong anticyclones are commonly associated with strong outbreaks of dense cold air that surge from the arctic to lower latitudes. Central United States and the Gulf of Mexico often experience this effect during the cold months of the year.

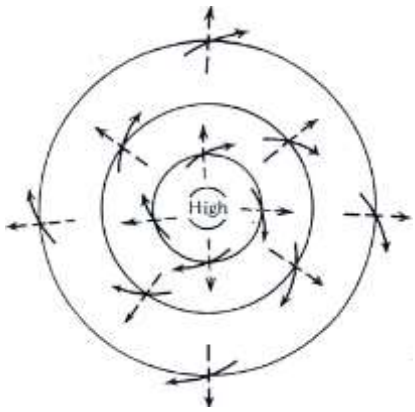


Fig 4-7: Anticyclonic circulation in the Northern Hemisphere.

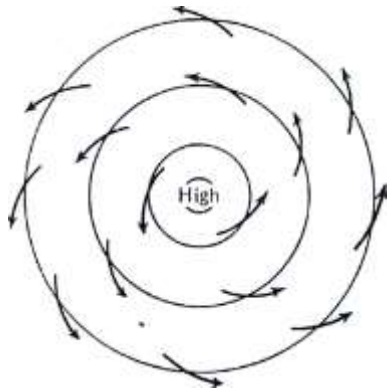


Fig4.8: Anticyclonic circulation in the Southern Hemisphere.

At such times, severe cold winds pour southward along the eastern half of an elongated anticyclone, bringing winds which may reach whole gale force to the interior United States and the Gulf of Mexico. Such strong cold winds are known as northers.

SESSION 5: FOG AND THUNDERSTORMS

5.1 Fog

Physically, there is very little difference between a fog and a cloud. They are both composed of minute water droplets suspended in the air. Fog, however, forms in the air near the earth's surface whereas clouds are features of much higher altitudes. Essentially, then, the difference between fog and clouds is one of method and place of formation rather than of structure or appearance.

Clouds form when the air cools adiabatically through rising and expanding. Fog forms through cooling of the air by contact and mixing, or on occasions through saturation of the air by increasing the water content. Frequently a continuous gradation exists from thick fogs into low-lying clouds, there being no definite distinction in appearance. The type of fog that forms depends on existing conditions and falls into four recognized categories: radiation fog, advection fog, frontal fog, and upslope fog.

In general, then, if surface air (the air near the earth's surface) is close to and is approaching the dew point (as determined by the psychrometer), fog formation can be anticipated. If the temperature should increase after the fog has formed, the dispersal of the fog may be expected. The thickness of the fog depends on various factors of humidity, temperature, wind, nuclei, etc.

Fogs are usually classified according to their effect on visibility. These conditions are given in Table 5-1.

TABLE 5-1 Visibility in Fog

	Objects Not Visible At
Dense fog	45 meters [50 yards]
Thick fog	180 meters [200 yards]
Fog	450 meters [500 yards]
Moderate fog	900 meters [V2 nautical mile]
Thin fog	1,800 meters [1 nautical mile]

5.2 Fog and Types of Fog

Fog is often described as a stratus cloud resting near the ground. Fog forms when the temperature and dew point of the air approach the same value (i.e., dew-point spread is less than 5°F) either through cooling of the air (producing advection, radiation, or upslope fog) or by adding enough moisture to raise the dew point (producing steam or frontal fog). When composed of ice crystals, it is called ice fog.

(1) Advection fog: Advection fog forms due to moist air moving over a colder surface, and the resulting cooling of the near-surface air to below its dew-point temperature. Advection fog occurs over both water (e.g., steam fog) and land.

(2) Radiation fog (ground or valley fog): Radiational cooling produces this type of fog. Under stable nighttime conditions, long-wave radiation is emitted by the ground; this cools the ground, which causes a temperature inversion. In turn, moist air near the ground cools to its dew point. Depending upon ground moisture content, moisture may evaporate into the air, raising the dew point of this stable layer, accelerating radiation fog formation.

(3) Upslope fog (Cheyenne fog): This type occurs when sloping terrain lifts air, cooling it adiabatically to its dew point and saturation. Upslope fog may be viewed as either a stratus cloud or fog, depending on the point of reference of the observer. Upslope fog generally forms at the higher elevations and builds downward into valleys. This fog can maintain itself at higher wind speeds because of increased lift and adiabatic cooling. Upslope winds more than 10 to 12 knots usually result in stratus rather than fog. The east slope of the Rocky Mountains is a prime location for this type of fog.

(4) Steam fog (arctic sea smoke): In northern latitudes, steam fog forms when water vapor is added to air that is much colder, then condenses into fog. It is commonly seen as wisps of vapor emanating from the surface of water. This fog is most common in middle latitudes near lakes and rivers during autumn and early winter, when waters are still warm and colder air masses prevail. A strong inversion confines the upward mixing to a relatively shallow layer within which the fog collects and assumes a uniform density. Under these conditions, the visibility is often 3/16 mile (300 meters) or less.

(5) Frontal fog: Associated with frontal zones and frontal passages, this type of fog can be divided into three types: warm-front pre-frontal fog; cold front post-frontal fog; and frontal-passage fog. Pre-frontal and post-frontal fog are caused by rain falling into cold stable air thus raising the dew point. Frontal passage fog can occur in a number of situations: when warm and cold air masses, each near saturation, are mixed by very light winds in the frontal zone; when relatively warm air is suddenly cooled over moist ground with the passage of a well-marked precipitation cold front; and in low-latitude summer, where evaporation of frontal-passage rain water cools the surface and overlying air and adds sufficient moisture to form fog.

(6) Ice fog: Ice fog is composed of ice crystals instead of water droplets and forms in extremely cold, arctic air (−29°C (−20°F) and colder). Ice fog of significant density is found near human habitation, in extremely cold air, and where burning of hydrocarbon fuels adds large quantities of water vapor to the air. Steam vents, motor vehicle exhausts, and jet exhausts are major sources of water vapor that produce ice fog. A strong low-level inversion contributes to ice fog formation by trapping and concentrating the moisture in a shallow layer.

In summary, the following characteristics are important to consider when forecasting fog:

- Synoptic situation, time of year, and station climatology.
- Thermal (static) stability of the air, amount of air cooling and moistening expected, wind strength, and dew-point depression.
- Trajectory of the air over types of underlying surfaces (i.e., cooler surfaces or bodies of water).
- Terrain, topography, and land surface characteristics.

5.3 Fog Characteristics

A general summary of characteristics important to fog formation and dissipation are given here. This is followed by general fog forecasting guidance and guidance specific to advection, radiation, and frontal fogs.

Formation: Fog forms by increasing moisture and/or cooling the air. Moisture is increased by the following:

- Precipitation.
- Evaporation from wet surfaces.
- Moisture advection. Cooling of the air results from the following:
- Radiational cooling.
- Advection over a cold surface.
- Upslope flow.
- Evaporation.

Dissipation: Removing moisture and/or heating the air dissipates fog and stratus. Moisture is decreased by the following:

- Turbulent transfer of moisture downward to the surface (e.g., to form dew or frost).
- Turbulent mixing of the fog layer with adjacent drier air.
- Advection of drier air.
- Condensation of the water vapor to clouds. Heating of the air results from the following:
- Turbulent transport of heat upward from air in contact with warm ground.
- Advection of warmer air.
- Transport of the air over a warmer land surface.
- Adiabatic warming of the air through subsidence or downslope motion.
- Turbulent mixing of the fog layer with adjacent warmer air aloft.
- Release of latent heat associated with the formation of clouds.

General Forecasting Guidance: In general, the following has been considered-

- Fog may thin after sunrise when the lapse rate becomes moist adiabatic in the first few hundred feet above ground.
- Fog lifts to stratus when the lapse rate approaches dry adiabatic.
- Marked downslope flow prevents fog formation.

- The moister the ground, the higher the probability of fog formation.
- Atmospheric moisture tends to sublimate on snow, making fog formation less likely.
- Rapid formation or clearing of clouds can be decisive in fog formation. Rapid clearing at night after precipitation is especially favorable for the formation of radiation fog.
- The wind speed forecast is important because speed decreases may lead to the formation of radiation fog. Conversely, increases can prevent fog, dissipate radiation fog, or increase the severity of advection fog.
- A combination advection-radiation fog is common at stations near warm water surfaces.
- In areas with high concentrations of atmospheric pollutants, condensation into fog can begin before the relative humidity reaches 100 percent.
- The visibility in fog depends on the amount of water vapor available to form droplets and on the size of the droplets formed. At locations with large amounts of combustion products in the air, dense fog can occur with a relatively small water vapor content.
- After sunrise, the faster the ground temperature rises, the faster fog and stratus clouds dissipate.
- Solar insolation often lifts radiation fog into thin multiple layers of stratus clouds.
- If solar heating persists and higher clouds do not block surface heating, radiation fog usually dissipates.
- Solar heating may lift advection fog into a single layer of stratus clouds and eventually dissipate the fog if the insolation is sufficiently strong.

Specific Forecasting Guidance: Consider the following when faced with advection, radiation, or frontal fog situations.

(a) Advection Fog: Advection fog is relatively shallow and accompanied by a surface-based inversion. The depth of this fog increases with increasing wind speed. Other favorable conditions include:

- Light winds, 3 to 9 knots. Greater turbulent mixing associated with wind speeds more than 9 knots usually cause advection fog to lift into a low stratus cloud deck.
- Coastal areas where moist air is advected over water cooled by upwelling. During late afternoon, such fog banks may be advected inland by sea breezes or changing synoptic flow. These fogs usually dissipate over warmer land; if they persist through late afternoon, they can advent well inland after evening cooling and last until convection develops the following morning.
- In winter when warm, moist air flows over colder land. This is commonly seen over the southern or central United States and the coastal areas of Korea and Europe. Because the ground often cools by radiation cooling, fog in these areas is called advection-radiation fog, a combination of radiation and advection fogs.
- Warm, moist air that is cooled to saturation as it moves over cold-water forms sea fog:
- If the initial dew point is less than the coldest water temperature, sea fog formation is unlikely. In poleward-moving air, or in air that has previously traversed a warm ocean current, the dew point is usually higher than the cold-water temperature.
- Sea fog dissipates if a change in wind direction carries the fog over a warmer surface.

- An increase in the wind speed can temporarily raise a surface fog into a stratus deck. Over very cold water, dense sea fog may persist even with high winds.
- The movement of sea fog onshore to warmer land leads to rapid dissipation. With heating from below, the fog lifts, forming a stratus deck. With further heating, this stratus layer changes into a stratocumulus cloud layer and eventually into convective clouds or dissipates entirely.

(b) Radiation Fog: Radiation fog occurs in air with a high dew point. This condition ensures radiation cooling lowers the air temperature to the dew point. The first step in making a good radiation fog forecast is to accurately predict the nighttime minimum temperature. Additional factors include the following:

- Air near the ground becomes saturated. When the ground surface is dry in the early evening, the dewpoint temperature of the air may drop slightly during the night due to condensation of some water vapor as dew or frost.
- In calm conditions, this type of fog is limited to a shallow layer near the ground; wind speeds of 2-7 knots bring more moist air in contact with the cool surface and cause the fog layer to thicken. A stronger breeze prevents formation of radiation fog due to mixing with drier air aloft.
- Constant or increasing dew points with height in the lowest 200 to 300 feet, so that slight mixing increases the humidity.
- Stable air mass with cloud cover during the day, clear skies at night, light winds, and moist air near the surface. These conditions often occur with a stationary, high pressure area.
- Relatively long time for radiational cooling, e.g., long nights and short days associated with late fall and winter in humid climates of the middle latitudes.
- In nearly saturated air, light rainfall will trigger the formation of ground fog.
- In valleys, radiation fog formation is enhanced due to cooling from cold air drainage. This cooled air can result in very dense fog.
- In hilly or mountainous areas, an upper level type of radiation fog—continental high inversion fog— forms in the winter with moist air underlying a subsiding anticyclone:
- Often a stratus deck forms at the base of the subsidence inversion and lowers. Since the subsiding air above the inversion is relatively clean and dry, air at the top of the cloud deck cools by long-wave radiational cooling which intensifies the inversion and thickens the stratus layer.
- A persistent form of continental high-inversion fog occurs in valleys affected by maritime polar air. The moist maritime air may become trapped in these valleys beneath a subsiding stagnant high-pressure cell for periods of two weeks or longer. Nocturnal long wave radiational cooling of the maritime air in the valley causes stratus clouds to form for a few hours the first night after the air becomes trapped. These stratus clouds usually dissipate with surface heating the following day. On each successive night, the stratus cloud deck thickens and lasts longer into the next day. The presence of fallen snow adds moisture and reduces daytime warming, further intensifying the stratus and fog. In the absence of airmass changes, eventually the stratus clouds lower to the ground.

- The first indicator of formation of persistent high-inversion fog is the presence of a well-established, stagnant high-pressure system at the surface and 700-mb level. In addition, a strong subsidence inversion separates very humid air from a dry air mass aloft over the area of interest. The weakening or movement of the high-pressure system and the approach of a surface front dissipates this type of fog.
- Radiation fog sometimes forms about 100 feet (30 meters) above ground and builds downward. When this happens, surface temperature rises sharply. Similarly, an unexpected rise in surface temperature can indicate impending deterioration of visibility and ceiling due to fog.
- Finally, radiation fog dissipates from the edges toward the center. This area is not a favorable area for cumulus or thunderstorm development.

(c) Frontal fog: Frontal fog forms from the evaporation of warm precipitation as it falls into drier, colder air in a frontal system.

- Pre-frontal, or warm-frontal, fog (Figure -8) is the most common and often occurs over widespread areas ahead of warm fronts.
- Whenever the rain temperature exceeds the wet-bulb temperature of the cold air, fog or stratus form.
- Fog usually dissipates after frontal passage due to increasing temperatures and surface winds.
- Post-frontal, or cold-frontal, fog occurs less frequently than warm-frontal fog.
- Slow-moving, shallow-sloped cold fronts (Figure-9), characterized by vertically decreasing winds through the frontal surface, produce persistent, widespread areas of fog and stratus clouds 150 to 250 miles behind the surface frontal position to at least the intersection of the frontal boundary with the 850 mb.
- Strong turbulent mixing behind fast moving cold fronts, characterized by vertically increasing winds through the frontal surface, often produce stratus clouds but no fog.

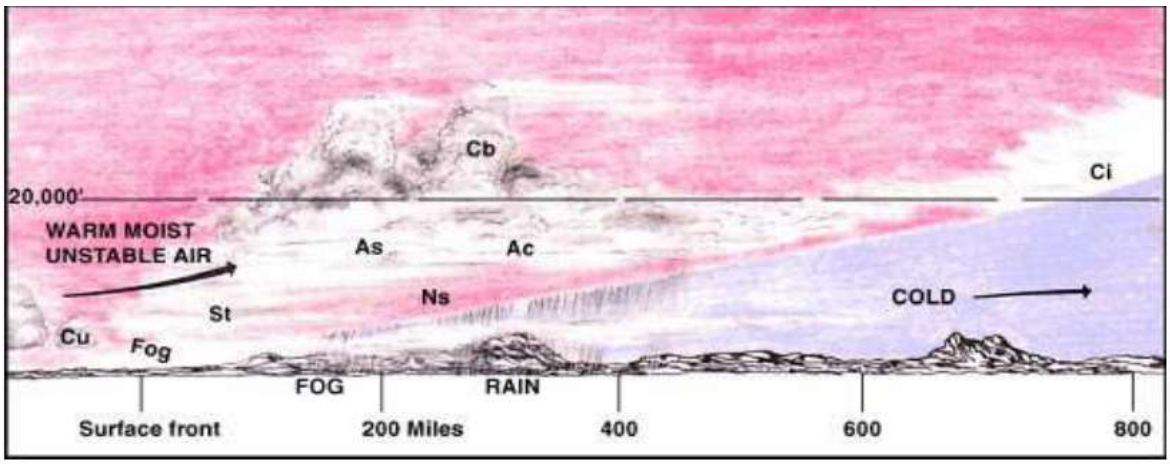


Figure-8: Pre-frntal Fog Associated with Warm Fronts

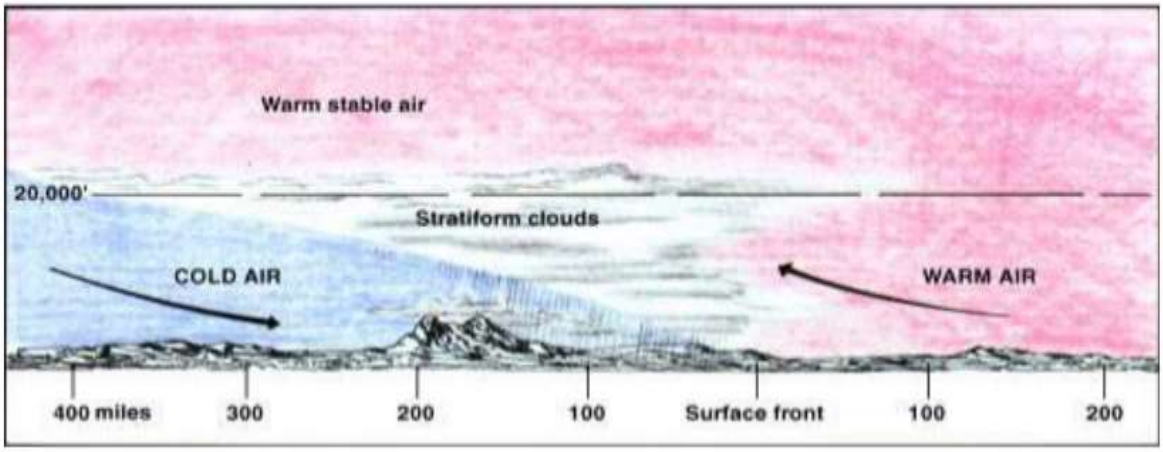


Figure-9: Post- frontal Fog Associated with Slow-Moving Cold Fronts

Radiation Fog: Radiation fogs are less widespread than the other types. They form under the

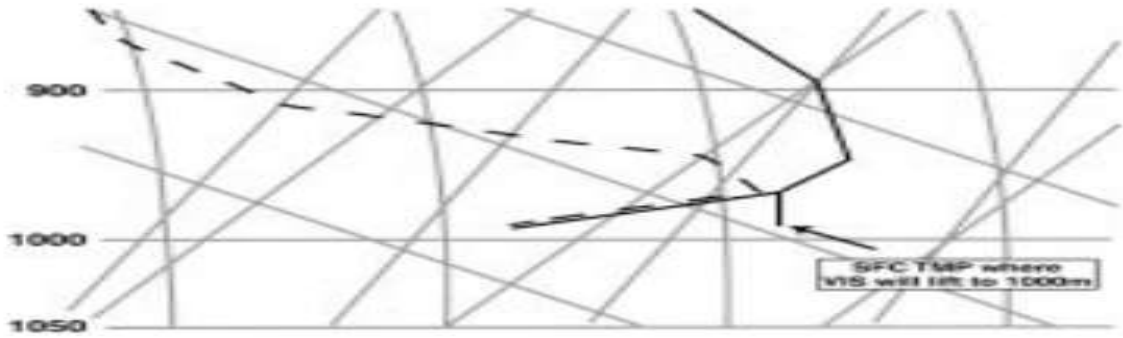


Figure-10: Radiation Fog

same conditions, though further advanced, that form dew and frost. On clear nights, with very slight wind, the earth and consequently the air above it will cool rapidly. If the air is cooled to its dew point to a greater depth than the slight layer necessary for dew formation, the resulting condensation will form not only on the ground but also on the nuclei of condensation in the air. This yields minute droplets suspended in the air which constitute fog. Since the cooling of the air depends on the earth beneath cooling rapidly by radiation, it is known as radiation fog.

Obviously, there will always be dew or frost when a radiation fog forms. Since clear nights are always necessary, radiation fog, in addition to do or frost, usually indicates fair weather for 12 hours at least. A slight wind is necessary in order to stir the cold air in contact with the ground and scatter it sufficiently so that fog forms. Further, the fog forms first and is thickest in the low areas or depressions since cool, relatively heavy air flows to the lowest points. Thus, radiation fog is sometimes known as ground fog owing to this tendency to "hug" the ground. Harbors, especially those surrounded by forested hills, are often shrouded in radiation fog during the night and early morning. Then, as the sun comes up over the horizon the fog is said

to "lift" of "burn off." Here again we have an excellent example of the earth's influence in heating the air. The fog is not simply blown away and does not actually rise. Rather, it evaporates as the air is warmed. But does it evaporate from the top down? On the contrary, the rising sun warms the earth, which then warms the surface air first. Hence the fog evaporates from the bottom up, or lifts. The last remnant of a thick radiation fog may therefore look like a low white cloud extending outward from the hilltops or other high spots.

In certain areas where inversions are frequent, the cold, heavier air is trapped beneath a lid of overlying warm air. Consequently, during nights favorable to radiation, pronounced cooling occurs in this cold air layer which may have considerable thickness and yield thick, high fogs that dispel slowly even after the sun rises.

5.4 Thunderstorms

Thunderstorms are among the most violent displays of nature. The violent and gusty winds which accompany the storm are deadly for small surface craft, and the extremely severe turbulence within the storm cloud is even more deadly for aircraft. All of the storm features—heavy showers, hail, severe gusty winds, and frequent lightning and thunder are the products of a rather localized giant convection cell in the atmosphere. The most striking visible result of this convection is the towering dark cumulonimbus cloud which usually begins as a simple cumulus cloud. The combination of adequate water vapor and steeply ascending air produces a strong vertical development of the original cumulus into the billowing cumulonimbus cloud whose top may be miles above its base.

Thunderstorm Development and Structure: Although the development of a thunderstorm is a continuous process, the events in this genesis can be grouped conveniently into three stages: the cumulus stage, the mature stage, and the dissipation stage. The most complete study so far has been that of the Thunderstorm Project of the National Weather Service, U. S. Navy, U. S. Air Force, and the former National Advisory Committee for Aeronautics (now the National Aeronautics and Space Administration). Although the three stages to be described are based on observations of Florida storms, they are probably applicable to most thunderstorms.

The cumulus stage is characterized by updrafts throughout the cell which become stronger toward the upper portion of the cloud where speeds up to 35 miles per hour [16 meters per second] occur. Although much weaker, the updraft actually begins at the ground. Air also enters the cumulus from the sides of the cloud, this process being called entrainment. This stage may be considered to begin when the cumulus grows above the freezing level, after which water and ice particles grow large enough to give a radar echo. The cloud remains in the cumulus stage for about 15 minutes, during which interval it can grow to 25,000 of 30,000 feet (somewhat lower in high latitudes).

The mature stage begins with the initial appearance of rain at the surface after drops have grown beyond the size supportable by the updrafts. Updraft speed still increases with elevation and appears to be at a maximum in the early mature stage when speeds of 70 miles per hour may occur, while strongest downward speeds are a little more than half this. The region of principal organized downdrafts being initiated by the drag of falling water and ice particles, together with the effects of cool air entrainment from the sides. When the downdraft reaches the surface, the air spreads horizontally, with the strongest motion being in the direction of storm travel—usually eastward. Associated with the downdraft are the heaviest showers of the storm, considerable gustiness, a fall in temperature, and an abrupt increase in pressure—all of which are experienced by a ground or sea-level observer. Hail occurs during the mature stage in many but not all storms and grows in concentric zones from particles being carried cyclically above and below the freezing level. During the mature stage, about 15 to 30 minutes, the thunderstorm cell reaches its maximum vertical extent— 30,000 to 60,000 feet. Upper westerlies may flatten out the upper portion of the cloud into the well-known anvil top, from which thin pseudo cirrus may be blown.

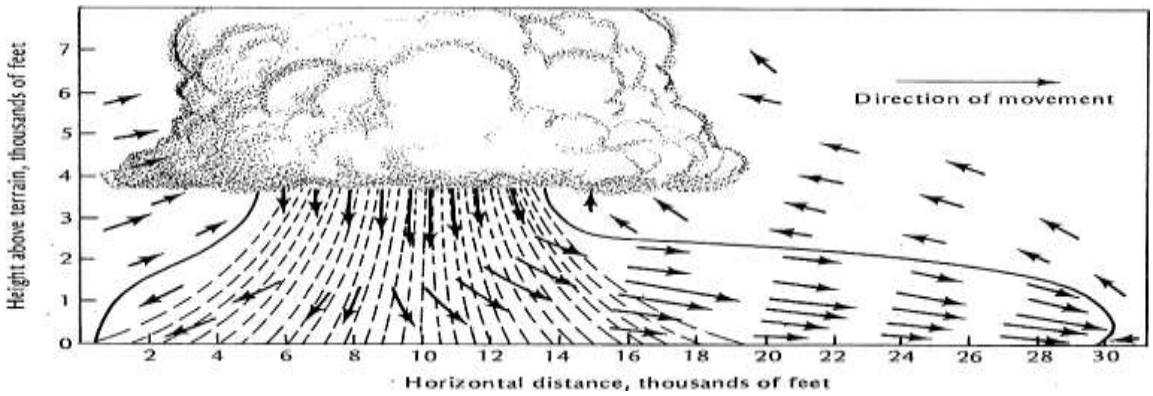


Fig. 5.1 *Air motion and precipitation beneath a cumulonimbus cloud,* (after H. Byers and R. Braham, The Thunderstorm, U.S. Government Printing Office)

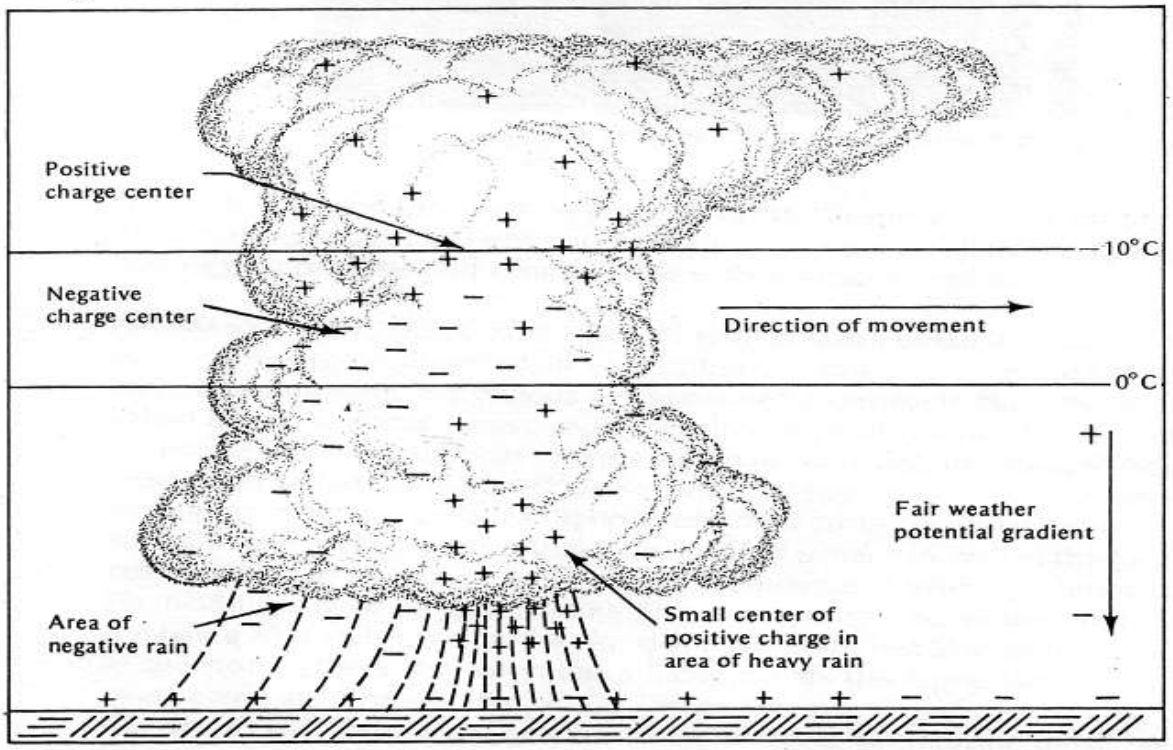
The dissipation stage begins when the region of downdraft has spread over the entire cell in the lower levels. During this stage the downdraft region spreads vertically as well as horizontally until the entire cell becomes a region of either weak downdraft or air with no motion. The structure and development just given refer to a single thundercloud. However, it is now known that many thunderstorms actually consist of a group or cluster of individual cells in successive stages of development. New cells develop on the forward, or strong downdraft, side of more mature cells.

Lightning and Thunder: In fair weather the earth normally has a negative charge with respect to the air, the potential gradient being directed downward at approximately 100 volts per meter. The intense friction of air on hydrometeors within the cumulonimbus builds very high charges in such clouds, with a clustering of positive charges in the upper portion and negative charges in the lower portion of the cloud; the latter results in a positive charge in the surface beneath, reversing the fair-weather occurrence. When the potential difference becomes high enough, discharges (lightning) may occur from cloud to ground, cloud to cloud, or within the

same cloud. The rapid expansion and contraction of highly heated air along the discharge path produces the almost explosive sound called thunder, which may reverberate by successive reflection from neighboring clouds. Sheet lightning is simply the general illumination of the sky produced by a lightning streak obscured by clouds or below the horizon.

Frequently in thundery weather at sea, a high voltage or potential difference is built up between a ship and the air or clouds above. This concentration of static electricity on the vessel tends to leap from all the pointed objects such as the ends of masts and spars. A purplish or bluish spray of light results, called St. Elmo's fire. Balls of lightning have been observed rolling along the masts and rigging of ships and disappearing with a sudden explosion.

Fig. 6-17 (a) Distribution of electric charges within and below a cumulonimbus cloud.



Air-Mass Thunderstorms: We will note in more detail in Chap. 11 that air masses are huge volumes of air with rather uniform characteristics of temperature and humidity. Air masses are separated from each other by "surfaces" called frontal surfaces whose characteristics depend on the air-mass motion.

Thunderstorms that occur entirely within an air mass, unrelated to frontal effects, are called air-mass thunderstorms. Many such storms originate as isolated convection cells. On land these tend to occur in late afternoons in summer when local overheating of the ground surface increases the local lapse rate so much as to cause instability. This sequence of events is summarized. In (a) local heating produces a simple cumulus cloud in which the rising air is stable within the cloud and comes to rest at a relatively low level at the point where the saturated adiabat meets the lapse rate for the non-overheated air. The situation in (b) shows that continued overheating causes the dry air rising from the ground to be unstable as the local lapse rate now exceeds the adiabatic rate. Within the cloud the saturated air is still unstable as the lapse-rate curve exceeds the moist adiabatic lapse rate.

Fig: Lightning strokes hitting the Empire State Building

Local thunderstorms of marine origin are most frequent in the early morning hours. On a clear night the air temperature may fall considerably at high levels, while the sea itself, as explained earlier, retains the heat of the day for long periods. Consequently, the air adjacent to the sea remains warm. During the night the temperature contrast between warm sea-level air and high-altitude air becomes more pronounced, developing a high lapse rate, necessary to all thunderstorms. Finally, any surface-temperature inequality causes a local rising air column to form, culminating in a storm.

Over land, local thunderstorms are most common in the late afternoon, when the effect of solar heating of the land surface is greatest, although nocturnal thunderstorms are quite common in the Northern Plains states of the United States.

Air-mass thunderstorms also occur from advection effects. The lapse rate may be increased through the advection (horizontal movement) of warm air near the ground or cold air at high levels. In either case a high lapse rate favorable to thunderstorm development occurs. Another important cause of air- mass thunderstorms is wind convergence. Uplift in the zone of convergence may trigger uplift in air already unstable causing the generation of a thundercloud. Or air which is in a state of convective instability may then become unstable from uplift in a zone of convergence, resulting in thundercloud development. When convergence occurs along a line, a form known as a line thunderstorm may be generated in the rising air above the linear zone of convergence.

Frontal Thunderstorms: These constitute another important class of line thunderstorm and tend to form in a warm air mass forced aloft over a steep cold front or underrunning boundary of a cold-air mass. This effect may trigger the uplift of air to great height if the warm air is either unstable to begin with or is convectively unstable. Frontal storms may occur at any time

or season but are most common and intense in late afternoon in summer when overheating tends to make for, or increase, instability in the warm air ahead of a moving cold front.

Squall-Line Thunderstorms: These storms, another of the line type of storms, are perhaps the most intense of all and are frequently associated with tornadoes. They occur in the warm air parallel to and preceding very vigorous cold fronts. Uplift in unstable warm air" ahead of an advancing front may be triggered by the effects of the front. Once generated, the line of storms then moves in the same direction as the front, but faster. Winds associated with these vigorous storms often exceed 60 miles per hour [27 meters per second] and cause considerable damage.

SESSION 6: TROPICAL CYCLONES

6.1 Tropical Cyclones

Tropical cyclones are the most violent storms experienced by the mariner. In West Indian waters these storms are known as hurricanes; in the East Indian and Japanese waters they are called typhoons; in the Indian Ocean they are called cyclones; off Australia, willy-willies; and off the Philippines, baguios. Technically they are tropical cyclones. Owing to common American usage, we shall use the terms hurricane and tropical cyclone interchangeably.

Although tropical cyclones are not regular features of the low latitudes, storms resembling them but of lesser intensity are much less rare. By international agreement this group of related storms has been classified according to intensity as follows: tropical depression—winds up to 34 knots [39 miles per hour]; tropical storm—winds of 35 to 63 knots [40 to 72 miles per hour]; hurricane or typhoon—winds of 64 knots [73 miles per hour] or higher.

A somewhat different classification that includes four storm divisions is widely used in the United States, as follows: tropical disturbance—shows a slight surface circulation with one or no closed isobars; tropical depression— has one or more closed isobars with wind equal to or less than 27 knots [31 miles per hour]; tropical storm—wind from 28 to 63 knots [32 to 72 miles per hour]; hurricane—wind speed greater than 63 knots [72 miles per hour].

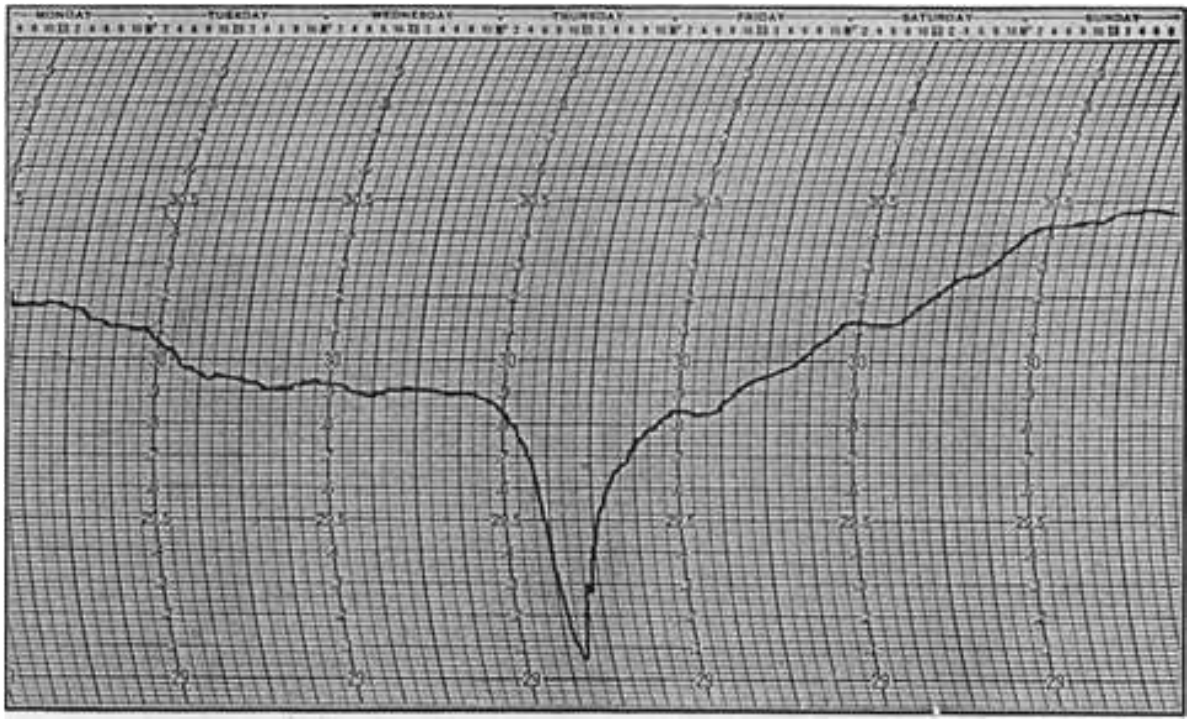
Tropical cyclones are relatively small, intense, low-pressure areas having a more or less circular shape. These storms are characterized by the circulation of a single air mass—tropical marine in character without fronts in contrast to the two-fold air-mass structure of mid-latitude extratropical cyclones which has a typical polar-front wave. Hurricanes are also much smaller than extratropical cyclones. Their size is rarely much greater than 300 nautical miles in diameter, and the very intense part is often more restricted.

The pressure gradient within the small hurricane area is extremely steep (Fig. 6-1). The fall of pressure from the periphery to the center of the storm commonly varies between 20 and 70 millibars. A record low pressure for Atlantic hurricanes was the 26.35 inches [892 millibars] reported for the storm of September 2, 1935 in the Florida Keys. The record low was reported as 26.185 inches [887 millibars] near Luzon on August 18, 1927. The gradient resulting from the severe drop is not radially uniform but increases in steepness toward the center.

The extremely high winds of the hurricane, which are directly related to this steep gradient, reach their maximum intensity near the center, owing to the increasing pressure gradient. This effect is shown schematically in Fig. 6-2. Note that as the wind spirals in toward the center it suffers increased deflection and finally travels in a circular path, never quite reaching the

center. Thus, a central region about 16 kilometers in diameter remains relatively calm. On opposite sides of the calm center the winds blow in opposite directions. The abruptness with which the wind changes as the center crosses a station is shown very strikingly by the wind recording in Fig. 6-3. Central winds between 100 and 200 knots are not at all uncommon.

Fig. 6-1 Microbarograph trace showing the passage of a tropical cyclone at Kings Point, New York on September 14, 1944.



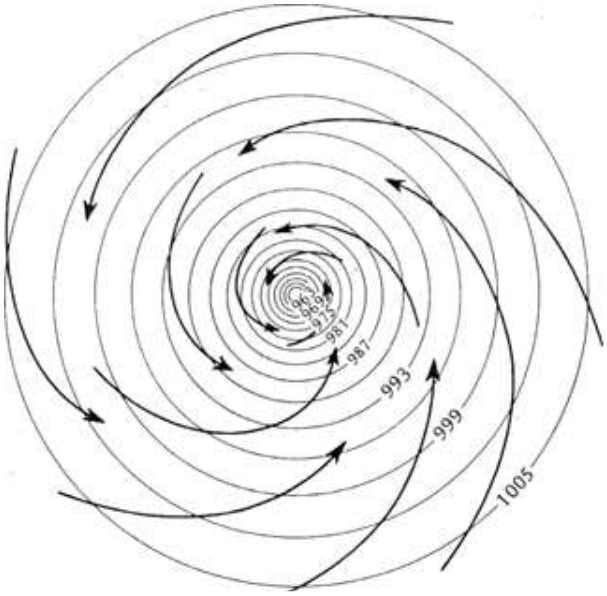


Fig. 6-2: increasing pressure gradient

It is often said that the winds set in with greater violence as the center passes over a vessel or station. The truth is that as the center is approached, the winds build up to maximum violence more or less slowly, compared to the change from dead calm to maximum velocity as the center passes. At the same time, the destructive effect of a sudden high wind is greater than that of a wind building up slowly to the same velocity. Thus, a line may snap under a sudden strain while it has withstood the same strain developed gradually. If the calm storm center passes across the vessel, preparations should always be made in anticipation of this sudden onslaught of the winds and their reversal. Many a vessel has keeled over and submerged following this sudden attack and reversal of the wind. The following quotation is from the report of the second officer on a ship passing through the hurricane of September 19, 1941, off western Mexico:

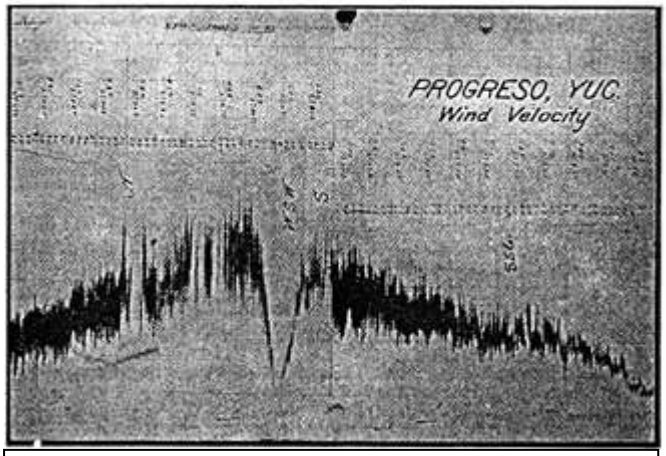


Fig. 6-3 Wind-velocity recording showing intensity and gustiness of winds preceding and following the calm center of a Yucatan hurricane, August 26, 1938. (Monthly Weather Review)

The barometer fell so rapidly that it could be seen moving down—from 29.25 to 27.67. At 4:30 the ship passed through the center of the cyclone. The wind died down to almost 0 and the low clouds opened up so that high cirrus could be seen through a small opening. There was a peculiar yellow light and the sea became bright green in color. The extremely low atmospheric pressure caused discomfort in the ears. High confused swells broke aboard the ship with terrific force from all sides. In about 10 to 15 minutes the center passed, and the wind came from the southwest, force 12 and over.

6.2 Vertical Structure: Clouds and Precipitation

As the wind spirals into the low-pressure storm center, the air rises as well. This rising motion is most nearly vertical surrounding the storm center. Rising air always cools, and in the humid marine air this cooling results in condensation in the form of clouds. These clouds vary from thin cirrostratus in the outer portions of the storm through altostratus and nimbostratus to cumulonimbus, where the steep, almost vertically rising air is encountered around the storm center. The calm center, having no generally rising air and often descending air, will be relatively clear. Thus, this calm center has a clear opening or shaft penetrating the surrounding heavy clouds and is called the eye of the storm (Fig. 6-4 and 6-7). The surrounding dense, black, cumulonimbus mass which rings the calm center is often called the bar of the storm, since it appears as a dense cloud bank when viewed from a distance.

This air motion responsible for the cloud formations. It should be noted that the tops of the central cumulonimbus clouds produce cirrus at their peaks. These cirrus clouds blow outward from the storm center in all directions. Being masked by the storm clouds, the cirrus forms are not visible until the margin of the hurricane is reached. Rain falls in the inner central areas of the storm, becomes very heavy in the ring surrounding the center, and is often attended by lightning and thunder in this area. The intensity of the rain in the region surrounding the eye is usually torrential. Measurements of 15 to 30 centimeters [7 to 12 inches] in 24 hours are outrun the storm. While under the influence of the wind in the storm the waves are irregular and often confused-looking. But after leaving or outrunning the storm, the waves become more and more regular and long crested as they slowly diminish in height and are then known as swell. The general way in which swells radiate after development in the different portions of the hurricane. Note that the maximum swell (generated in the right side) tends to move ahead of, but in the direction of, the storm and often is an early forerunner of a hurricane directly approaching an observer.

Acomparison between wind motion indicates that as waves move through the storm area, they must encounter winds that travel to the left of the wave direction, for the swell from upwind always deviates to the right of the existing wind. The amount of this deviation depends upon the speed and path of the storm.

6.3 The Storm Surge

By far the greatest damage and loss of life and property from hurricanes are due not to the wind or torrential rains but to the catastrophic flooding that usually accompanies a hurricane whose path crosses a coastal region, whether continent or island. This rise in sea level is known as the meteorological or storm surge.

In part, the surge results from the impounding of water against the coast by the wind. Then waves tens of feet high, superposed on this raised sea level, inflict additional alternating onslaughts of water which create tremendous havoc. In addition to the slow build-up of the wind-driven seas and the impulsive effects of high wind waves, many hurricanes exhibit a further rather abrupt surge of water that strikes the coast about the same time as the center does. This surge is thought to be an effect of resonance produced in the following manner. The low-pressure center of the hurricane causes the ocean to rise like an inverted barometer, the uplift depending on the pressure decrease. Under appropriate conditions of water depth, the central "mound" of water becomes amplified as it travels and strikes the coast as a wave as much as 12 feet higher than normal sea level. If all of the above effects are superimposed on the normal diurnal high tide, particularly a spring tide, the results are all the more disastrous.

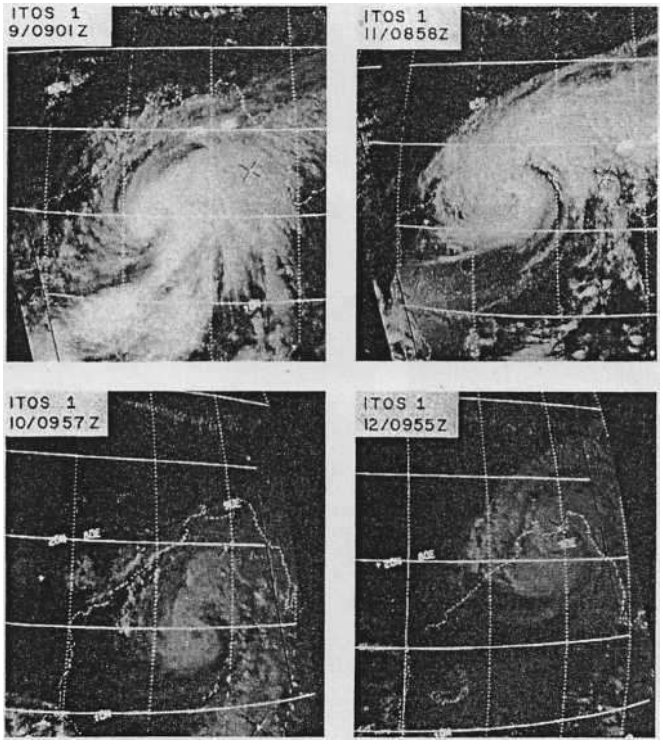
Examples of the devastation caused by the storm surge of a hurricane were never more strikingly shown than in the tropical cyclone of November 1970 and by the Gulf hurricane Camille of August 1969.

The Great Cyclonic Storm of 1970: This was possibly one of the most devastating storms in history. When the storm-driven waters receded after passage of this Indian Ocean cyclone, an estimated 300,000 persons were left dead, mostly by drowning. The storm moved northward out of the Indian Ocean, as is evident in the satellite. The numerous low-lying islands off the coast of Bangladesh, as well as the shore region itself, were rather quickly submerged beneath waters that reached 4 meters above expected high-tide levels.

Flooding and devastation were contributed to by the pronounced regional tide range, which is from 3 to 6 meters. Even a small surge of water the time of high tide is extremely serious in view of the low relief of the islands and coast.

In addition to the immediate loss of life, the fishing industry, the major protein source for the region, suffered about a 65 percent loss through destruction of boats and shore facilities as well as through the loss of more than half of the fishers. Unfortunately, this region is beset with

frequent severe tropical cyclones whose deadliness is compounded by the high tides, funneling coastal configuration, and flat terrain.



Hurricane Camille, 1969: This was certainly one of the most powerful hurricanes ever to strike the shores of the United States. Camille came out of the Gulf of Mexico carrying wind gusts estimated at 190 miles per hour along the shore. The central pressure of 901 millibars [26.61 inches] was the second lowest recorded in this century. The storm tide that flooded the coast peaked at 20 feet above normal. About 200 persons lost their lives and damage was close to a half-billion dollars. From southern Mississippi the storm followed a sweeping arc across the Southeastern states to the Atlantic Ocean. Its power abated inland because the energy derived from condensed moisture lessened as the water supply decreased. But on August 19 and 20, Camille gained renewed moisture and energy from the moisture evaporated from the Atlantic Ocean. Torrential rains resulted in precipitation of up to 28 inches of rain in a 30-hour period over much of the high region of Virginia. Racing floodwaters, which took another 105 lives, left more than 100 million dollars of damage in their wake.

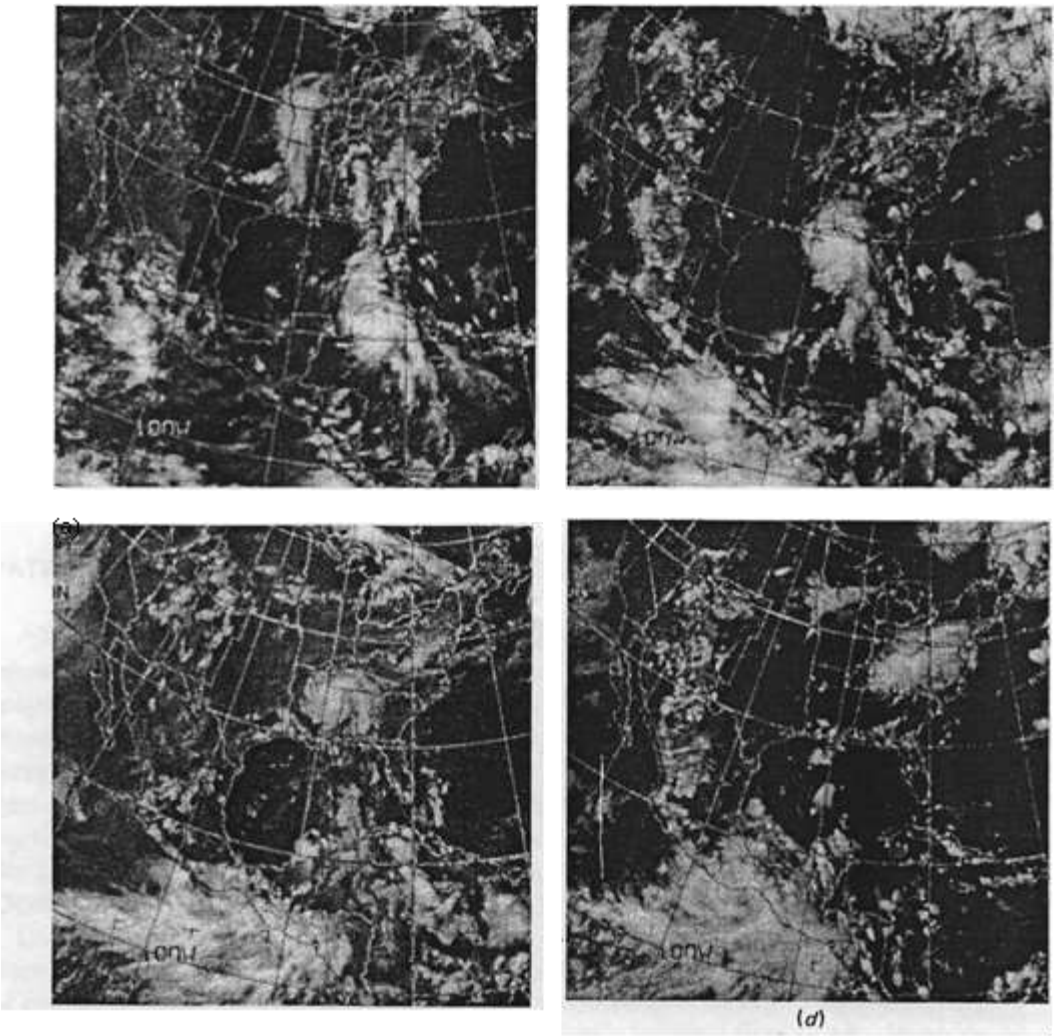


Fig. 6-5: Four successive views of hurricane Camille, August 1960. This powerful storm crossed the southern coast of Mississippi from the Gulf of Mexico and then cut a path across the Southeast-ern United States and back into the ocean.

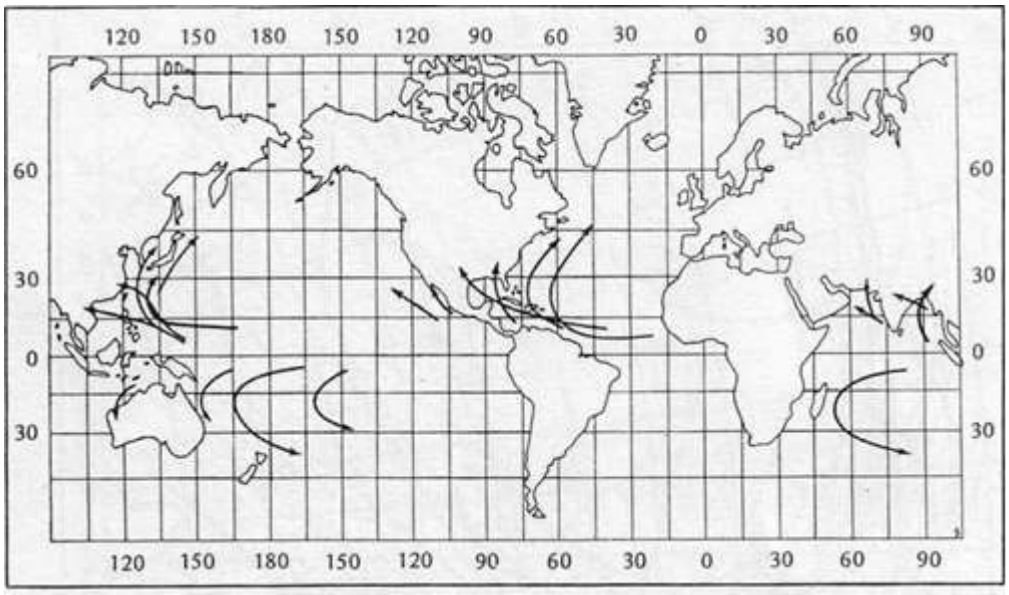


Fig. 6-6 The principal hurricane regions of the world and their average paths of motion

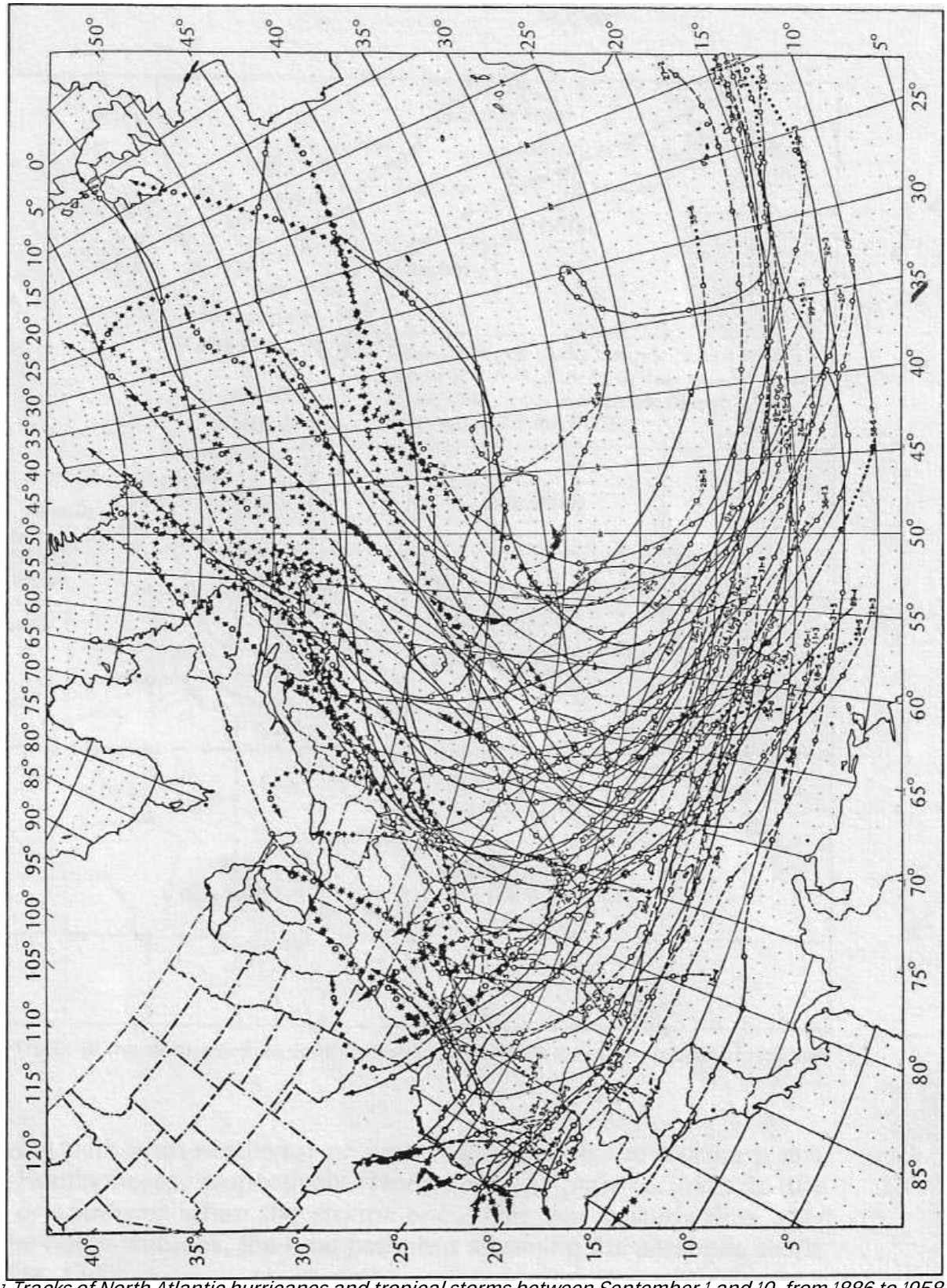
6.4 Paths and Regions of Occurrence

Although tropical cyclones often travel well into the middle latitudes, especially in the North Atlantic Ocean, they are essentially tropical and semi- tropical storms. The regions and average paths of motion of most of the tropical cyclones of the world. Fully developed tropical cyclones have never been reported from the South Atlantic nor from the South Pacific east of about 140° west longitude. The path and motion of an individual storm may depart considerably from the mean paths shown in the diagram. An example of individual paths over a fairly long interval for the North Atlantic Ocean, which illustrates well the diversity of individual storms.

Despite this diversity, some important generalities can be given. As is evident hurricanes begin near, but never on, the equator. The region of origin is usually from 6 to 10° away from the equator. At first, they move very slowly away from the equator but with the easterly tropical flow of air, developing a curving path west-northwest or west-southwest, in the Northern and Southern Hemispheres, respectively. Normally, this path recurves to the northeast or southeast when the storms encounter the westerly flow after crossing the horse latitudes, the total path thus assuming the parabolic shape.

Prior to recurvature, hurricanes have a relatively low forward speed—about 10 to 12 knots on the average. This motion accelerates upon recurvature to an average of 20 to 30 knots although individual storms have reached 60 knots, such as those along the Atlantic Coast in 1938 and 1944. They ultimately peter out or merge with extratropical wave cyclones and lose their identity.

Although hurricanes tend to follow the tracks influenced by the trades and then the westerlies, the paths of individual storms are also specifically controlled by the circulation aloft, which often involves the presence of strong low and high surface pressure cells whose effects reach high elevations. A strong high-pressure region, particularly north of the horse latitudes, may block the normal recurvature of a hurricane and deflect the storm to the westward, following the circulation about the anticyclone. Many West Indian hurricanes are thus directed northward along the eastern coast of the United States, around the western flank of a well-developed Bermuda high instead of recurving out to sea. The tracks of two such storms that wrought great havoc along the coast of the United States.



Tracks of North Atlantic hurricanes and tropical storms between September 1 and 10, from 1886 to 1958. Dotted portions represent tropical depressions for the development stage; dashed portions the tropical storm stage; solid lines the hurricane stage; crosses, extra tropical cyclone stage; and stars the dissipation stage. (National Weather Service)

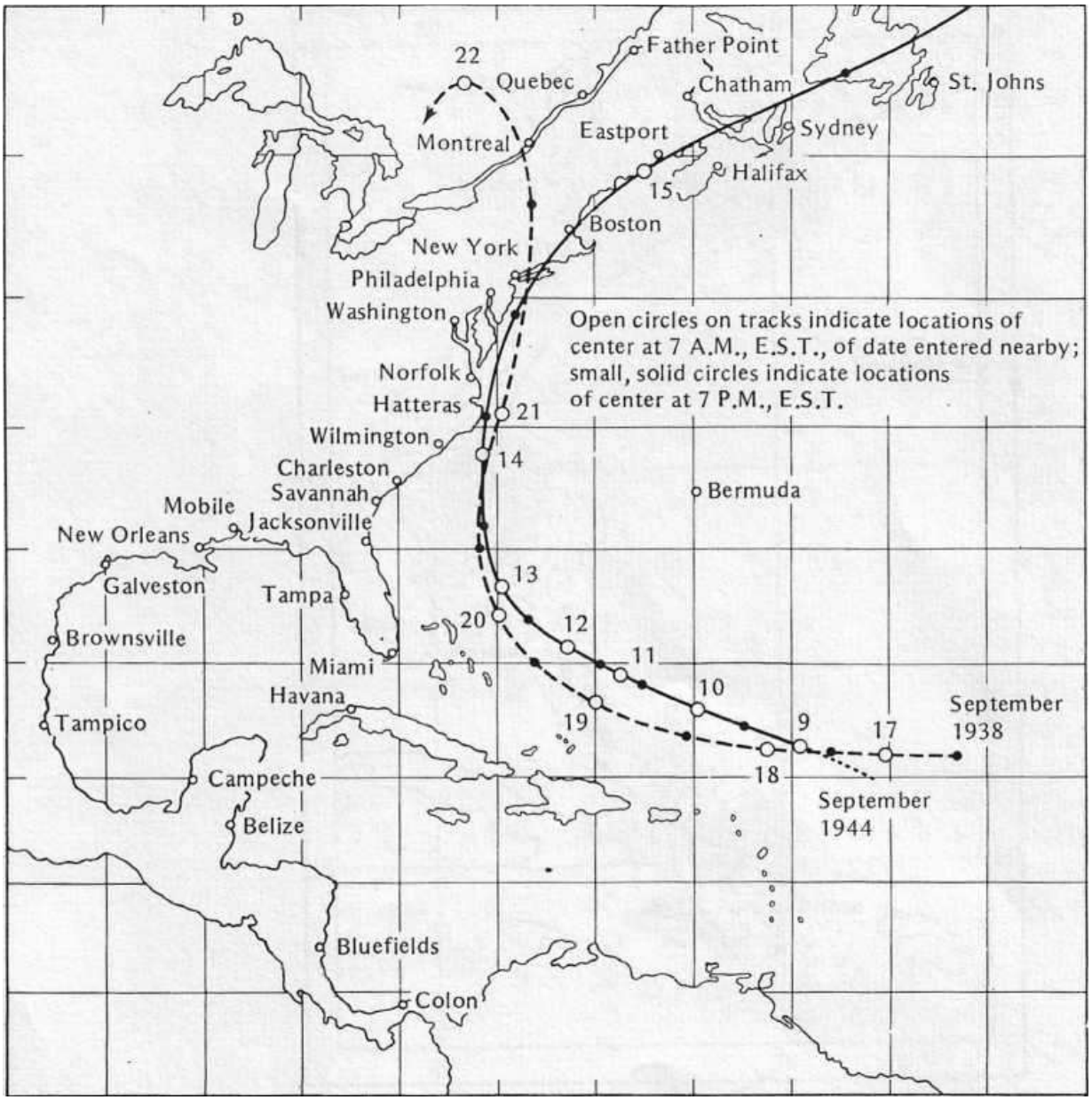


Fig. 6-8: Tracks of the destructive September hurricanes of 1938 and 1944 (National Weather Service)

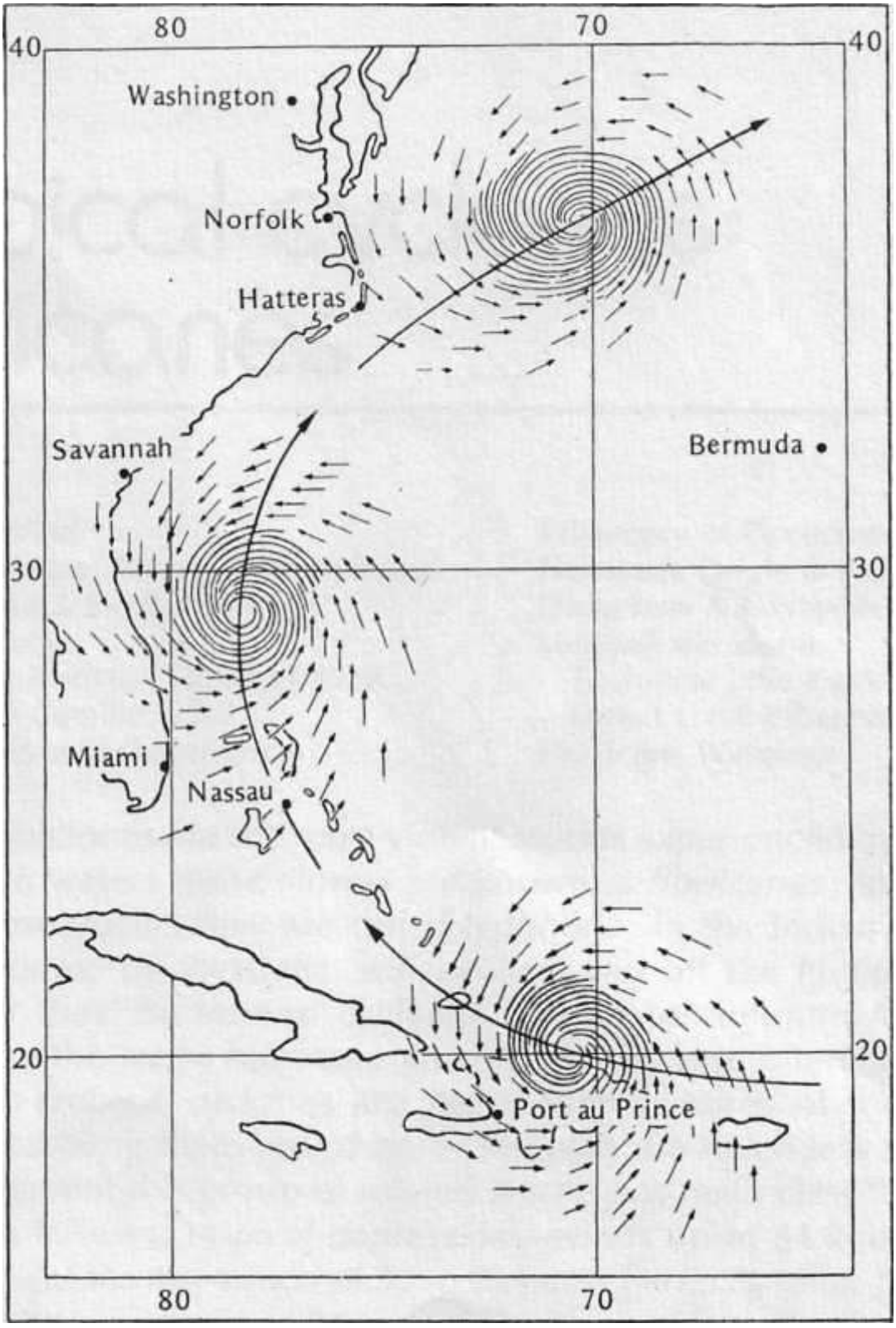


Fig. 6.9- Typical track and wind system of West Indian hurricanes, which commonly move north and then northeastward off the coast of North America.

SESSION 7: WEATHER FORECASTING

7.1 Weather Forecasting

Weather forecasting can be defined as the act of predicting future weather conditions or an attempt to indicate the weather conditions which are likely to occur.

Weather forecasting is the application of Science and Technology to predict the state of the atmosphere for a future time and a given location. Human beings have attempted to predict the weather informally for millennia, and formally since at least the nineteenth century. Weather forecasts are made by collecting qualitative data about the current state of the atmosphere and using scientific understanding of atmospheric processes to project how the atmosphere will evolve within the next few hours or more.

7.2 Methods of Weather Forecasting

The methods of weather forecasting are as follows:

- Persistence forecasting
- Climatology forecasting
- Looking at the sky
- Use of a barometer
- Nowcasting
- Use of Forecasting Models
- Analogue Forecasting
- Ensemble Forecasting

7.3 Applications of Weather Forecast

- Severe weather alerts and advisories
- Air Traffic
- Marine
- Agriculture
- Utility companies
- Private Sector

7.4 Factors Depend on Weather Forecasting

- the forecast range (medium-range, short-range, very short-range, nowcasting) and the size of the domain to be covered (large portion of the globe, regional domain, small country, city)
- the geographical context and related climatology (midlatitudes, tropical or equatorial areas, isolated islands)
- the potential risk associated with the expected weather at various ranges
- the organization of the forecast service (multipurpose forecasters or specialized forecasters for each type of applications)
- the end-user who receives the forecasts (civil defence, aviation, marine, hydrologic and water management service, road administration, medias, public)
- the technical environment (available external and/or internal NWP products, in situ observations, satellite and radar images, lightning detection network, efficient visualization workstation adapted to the forecaster, Web access).

7.5 Tasks Relevant to Weather Analysis and Forecasting

Weather analysis and forecasting should be considered as the succession of the following tasks:

- to clearly understand the recent evolution and the actual situation of the atmosphere at all time scales (weather analysis);
- to obtain the pertinent information from at least one numerical model and to assess the future evolution of the atmosphere in order to determine the most likely scenario (synoptic weather forecasting); when available, automated tools provide a first guess, that forecasters may or may not follow.
- to deduce finally the consequences of the expected synoptic situation in terms of weather elements (weather elements forecasting) and to evaluate the risk of the occurrence of hazardous phenomena (risk assessment).
- to prepare the meteorological information (including possible warnings) to be directed toward the various internal or external users.

7.6 Required Tasks to be Performed for Weather Forecasting

- Step 1: Evaluating the present meteorological situation
- Step 2: Examining the quality and relevance of the analysis
- Step 3: Identifying the key elements of the meteorological situation, according to the accepted conceptual models and / or guidance / tools
- Step 4: Examining the various guidance products and choosing the most likely scenario
- Step 5: Describing the evolution of the atmosphere corresponding to the chosen scenario
- Step 6: Deducing the consequences for smaller scales and specific areas

- Step 7: Describing of the expected weather in terms of weather elements (including automated production techniques when applicable)
- Step 8: Deciding on the opportunity / necessity to issue / end warnings
- Step 9: Distributing the various products to users
- Step 10: Evaluating according to performance measurements / Verifying forecasts.

7.7 Problems of Weather Forecasting

The problems of weather forecasting, as seen from the standpoints of mechanics and physics. If, as every scientifically inclined individual believes, atmospheric conditions develop according to natural laws from their precursors, it follows that the necessary and sufficient conditions for a rational solution of the problems of meteorological prediction are the following:

- The condition of the atmosphere must be known at a specific time with sufficient accuracy.
- The laws must be known, with sufficient accuracy, which determine the development of one weather condition from another.

7.8 Essentials of Weather Forecasting

Essential features of weather forecasting are:

- Proper recording of data.
- Careful study of synoptic charts.
- Search for similar situation from the historical data.
- Preparation of the weather condition chart as may be possible in next 24 hours, and
- Drawing quick, correct levels and definite conclusions regarding future weather phenomenon.

7.9 Elements included in Weather Forecasting

From another side, the elements of agricultural weather forecasts vary from place to place and from season to season, but they should refer to all weather elements, which affect farm planning and/or operations, and they ideally would include (WMO, 2001):

- Sky coverage by clouds
- Precipitation
- Temperature (maximum, minimum and dew point)
- Relative humidity
- Wind Speed and direction
- Extreme events (heat and cold waves fog, frost, hail, thunderstorms, wind squalls and gales, low pressure areas, different intensities of depressions, cyclones, tornados, …)
- Bright hours of sunshine

- Solar radiation
- Dew Leaf wetness
- Pan evaporation
- Soil moisture stress conditions and supplementary irrigation for rainfed crops
- Advice for irrigation timing and quantity in terms of pan evaporation
- Specific information about the evolution of meteorological variables into the canopy layer in some specific cases Micro-climate inside crops in specific cases.

7.10 Types of Weather Forecasting

Based on time or duration of forecasting period, the weather forecasting can be divided into six categories:

- Now-casting (NC)
- Very short-range weather forecasting
- Short range weather forecasting
- Medium range weather forecasting
- Extended range weather forecasting
- Long range weather forecasting

Now-Casting (NC): Current weather variables and 0-6 hour's description of forecasted weather variables. A relatively complete set of variables can be produced (air temperature and relative humidity, wind speed and direction, solar radiation, precipitation amount and type, cloud). Prerequisite is the operational continuity and the availability of an efficient broadcasting systems (e.g. very intense showers affecting a given territory must be followed with continuity in provision of information for final users). Accuracy is very high and potential usefulness is low.

Very Short-range Weather Forecasting: Up to 12 hours description of weather variables. A relatively complete set of variables can be produced (air temperature and relative humidity, wind speed and direction, solar radiation, precipitation amount and type, cloud). Prerequisite is the availability of an efficient broadcasting system (e.g. frost information must be broadcasted to farmers that can activate irrigation facilities or fires or other systems of protection). Accuracy is very high and potential usefulness is moderate.

Short Range Weather Forecasting (SRH): Short range weather forecasts are for a period of 12 hours to 72 hours. These daily forecasts are useful to irrigation engineers and farmers. A relatively complete set of variables can be produced (air temperature and relative humidity, wind speed and direction, solar radiation, precipitation amount and type, cloud). In SRF, the attention is centred on meso-scale features of different meteorological fields. SRF can be broadcasted by a wide set of media (newspapers, radio, TV, web, etc.) and can represent a fundamental information for farmers. Accuracy and potential usefulness are high.

Medium Range Weather Forecasting (MRF): Medium range weather forecasts are for periods of 3 to 10 days. A relatively complete set of variables can be produced (air temperature and

relative humidity, wind speed and direction, solar radiation, precipitation amount and type, cloud). In MRF the attention is centred on synoptic features of different meteorological fields. MRF can be broadcasted by a wide set of media (newspapers, radio, TV, web etc.) and can represent a fundamental information for farmers. Accuracy is high or moderate until 5 days; lower after and potential usefulness is very high.

Extended Range Weather Forecasting (ERF): Extended range weather forecasts are for periods of 10 to 30 days. Forecast is usually restricted to Temperature and precipitation.

Long Range Weather Forecasting (LRF): The long-range weather forecasts are issued thrice in year. Validity period of long-range weather forecast is 10 to 30 days. The long-range forecasts are useful for choosing cropping patterns.

SESSION 8: EARLY WARNING SYSTEM

Monitoring and warning services are what most often comes to mind when talking about EWS. This is the infrastructure that delivers forecasts and warnings. Monitoring is the act of collecting information along with a set of proxy variables related to risk, such as rain (correlated with floods/ droughts), or seismic waves (correlated with earthquakes). This can be done through direct observations, e.g. through seeing an approaching wildfire or landslide. However, such observations have limited usability due to the close proximity of the threat and limited possibilities for taking risk-reducing actions. The same problem arises with earthquakes, where seismometers are able to trigger a warning just seconds before impact. For many other hazards, where EWS allow warnings with a considerably longer lead time (weeks to months) such as storms, flooding, and droughts, continuous monitoring of significant proxy variables can trigger risk mitigating actions.

Modern technology provides the possibility of compiling data from multiple monitoring sources and at high speed, creating possibilities for improving the accuracy and One promising monitoring system is the emerging practice of sourcing data through social media. Numerous researchers have experimented with extrapolating from social media or from built-in smartphone sensors regarding nascent disasters. An example of a success in this area is represented by the I-React project (www.i-react.eu), which is led by the Italian research and innovation center 'Istituto Superiore Mario Boella' in partnership with different institutions and agencies (UNESCO among them). I-React is developing a strategy to upgrade risk management systems. In particular, I-React is developing new technologies such as a mobile app, wearable positioning systems, and a social media analysis tool to account for real-time crowdsourced information. This type of data collection can be especially helpful for gathering large datasets from the stakeholders personally affected by nascent disasters, while also adding location-specific knowledge to the provided information.

Modern technology provides the possibility of compiling data from multiple monitoring sources and at high speed, creating possibilities for improving the accuracy and rapidity with which disaster forecasts can be produced. Some common technological monitoring solutions that can provide such data are showcased below.

For many countries, historical weather information and monitoring infrastructure already exists, but may belong to different institutions, resulting in a fragmentation of monitoring capabilities as well as reducing the information available when making forecasts. If various stakeholders are able to bridge this gap, rapid improvements in forecasting rapidity abilities are possible. These arrangements can be achieved via legislation, institutional arrangements, or informal collaboration.

8.1 Introduction to Early Warning System

From 1991 to 2005, floods, windstorms, droughts, and landslides worldwide killed over 422,000 and affected over 3 billion people (International Strategy for Disaster Reduction 2006). In 2008, Cyclone Nargis devastated Myanmar, killing over 100,000 and displacing many others. Tropical cyclones during the 2000s, including Nargis, caused thousands of casualties, inflicted enormous economic losses, and caused considerable human suffering. Exposure to tropical cyclones has increased as more people have moved to vulnerable coastal locations than ever before. Likewise, vulnerability to wildfires has increased. In 2009, a series of large, rapidly moving bushfires devastated populated areas near Melbourne, Australia. Massive firefronts moved with incredible speed and ferocity, taking 173 lives and destroying thousands of homes. In 2010, a historic heat wave and numerous wildfires impacted Moscow and surrounding areas in the Russian Federation. At the same time, catastrophic flooding from unusually heavy monsoon rains was ongoing. Both events led to many fatalities and considerable human suffering. As climate change due to anthropogenic forcing continues, extreme weather events such as these are likely to become more common (IPCC 2007), further increasing the need for preparedness and early warning systems. The need for robust early warning systems goes beyond purely natural disasters and extends to include response to man-made disasters. In 2010, a catastrophic oil spill in the Gulf of Mexico devastated the ecosystem, and severely impacted the local fishing and tourism industries. In such a disaster, meteorologists must be prepared to work with emergency response officials and experts in other disciplines to mitigate the effects of the disaster through effective decision support services. The dramatic impact of natural disasters and the subsequent response activities often attract much international interest. Attention is being increasingly focused upon natural disasters inflicting tremendous economic losses (in addition to human suffering and casualties) and the efforts expended on the mitigation and reduction of such disasters. Disaster prevention and mitigation is now a recognized international priority. WMO

cooperates within many other organizations and international programs, particularly the International Strategy for Disaster Reduction (ISDR), in its efforts to improve natural disaster prevention and mitigation. In September 2006, Mr Kofi A. Annan, then the UN Secretary-General, said in the Forward to The Global Survey of Early Warning Systems, "Natural Hazards will always challenge us, but people-centered early warning systems can be a potent weapon in ensuring that natural hazards do not turn into unmanageable disasters." Increasingly, it is recognized that disasters are linked. The impacts of many types of natural disasters do not happen in isolation, but recognition of such cause and effect on a global and regional scale is leading to the creation of early warning systems that can accommodate multiple hazards and cross-boundary impacts. At the same time, governments are becoming aware that a paradigm shift from crisis management to risk management is necessary if the finite resources available are spent in the most efficient way to assist the populations at risk to prevent or mitigate disasters. The World Conference on Disaster Reduction, which was held at Hyogo, Japan in 2005, identified five (5) priority areas in the Framework of Action it adopted for 2005-2015. The second item on the list is "Identify, assess and monitor risks and enhance early warning." Natural hazards turn into disasters if the affected people cannot cope with them. A community without early warnings will be unprepared and suffer from the full-blown damages inflicted by the hazard. At its Fourteenth Session, the WMO Commission for Basic Systems (CBS, CBS-XIV, Dubrovnik, Croatia, 2009) requested the WMO Public Weather Services Programme (PWSP) to continue its focus on assisting Members to improve their national Public Weather Service programmes by providing guidance on the application of new technology and scientific research in data acquisition and use, especially for nowcasting and multi-hazard warnings. These guidelines are prepared with focus on the role of NMHSs in reducing the impact of disasters. The development of early warning systems is seen as part of the operational responsibility of NMHSs. The essential elements of such systems and in particular, forecasting, formatting, presenting and communicating of warnings of severe weather, and the accompanying public education and capacity building of NMHSs are given special attention in these guidelines. The application of nowcasting in warning operations and examples of nowcasting systems used by various NMHSs complete the document.

8.2 Framework of Risk Management

Wilhite et al. (2000) have taken the commonly accepted cycle of disaster management and redefined it in terms of crisis management and risk management. Crisis management emphasizes post disaster impact assessment, response, recovery, and

Schematic showing the inter-relationship of the elements of the Hyogo Framework for Actions

reconstruction, whereas risk management emphasizes protection through mitigation, preparedness, prediction, and early warning. Traditionally, the focus of disaster management has almost been exclusively on actions taken immediately before, during, and shortly after a disaster. The Hyogo Framework for Action 2005 – 2015 (HFA) (International Strategy for Disaster Reduction, 2005) provides the framework for a new paradigm in disaster risk management with a strong focus on prevention and preparedness strategies based on identification and quantification of potential risks. It encompasses risk identification, risk reduction and risk transfer. Figure 1 provides a simplified schematic of a comprehensive national strategy for disaster risk management derived from HFA.

There is no doubt that the role of relief assistance during a crisis will remain important and need to be enhanced at all levels. However, a paradigm shift is occurring with a move away from purely reactive response and recovery to a much more proactive and holistic concern about preparedness and prevention. Proactive mechanisms are sought to reduce the economic costs and impacts of hazards, improve response capacity, decrease vulnerability, and enhance communities' resilience to disasters.

8.3 Effective Early Warnings

The primary objective of a warning system is to empower individuals and communities to respond timely and appropriately to the hazards in order to reduce the risk of death, injury, property loss and damage. Warnings need to get the message across and stimulate those at risk to take action. Disaster mitigation decision makers require increasingly precise warnings to ensure effective measures may be formulated. Generally, demands for improvement in severe weather warnings take on the following form (Gunasekera 2004):

- extending the lead time of warnings.
- improving the accuracy of warnings.
- greater demand for probabilistic forecasts.
- better communication and dissemination of warnings.
- using new technologies to alert the public.
- targeting of the warning services to relevant and specific users (right information to right people at right time at the right place); and,
- warning messages are understood, and the appropriate action taken in response.

It should be noted that longer warning lead times should be considered together with the need to reduce the false alarm ratio and a balance should be achieved between the two whereby decisions can be based on optimum lead times for warnings. As described at the 3rd International Conference on Early Warning (EWC III, Bonn, Germany, 2006), effective early warning systems must be people-centered and must integrate four key elements:

- knowledge of the risks faced.
- technical monitoring and warning service.
- dissemination of meaningful warnings to those at risk; and,
- public awareness and preparedness to act.

8.4 Early Warning Systems

"Natural Hazards will always challenge us, but people-centered early warning systems can be a potent weapon in ensuring that natural hazards do not turn into unmanageable

8.4.1 Key Elements of Early Warning Systems

A complete and effective community based early warning system comprises four inter-related elements: risk knowledge, monitoring and warning service, dissemination and communication and response capability. A weakness or failure in any one part could result in failure of the whole system.

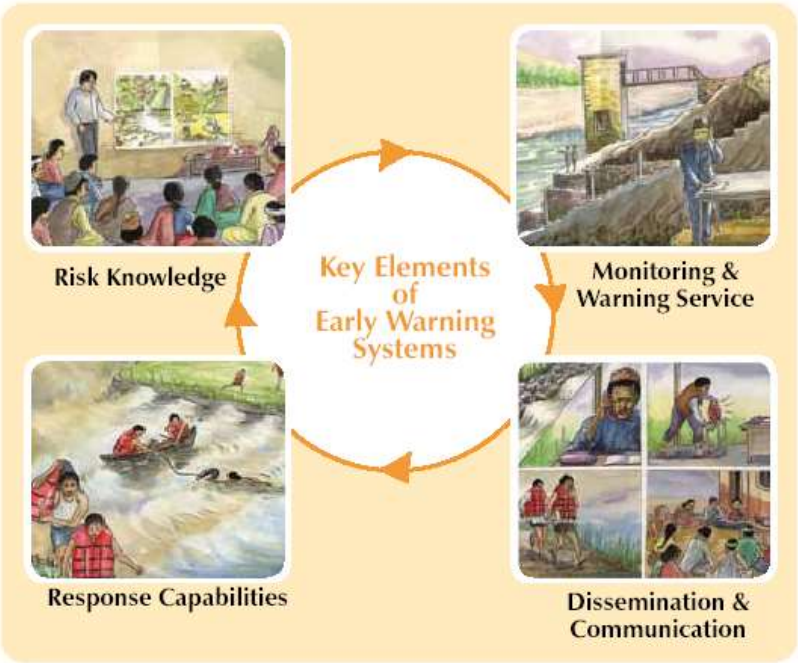


Fig. Early warning cycle

8.4.2. Essentials of EWS

Early Warning System/ Community based early warning systems must be built on four essential aspects:

a. Effectiveness
b. Efficiency
c. Equity and
d. Legitimacy

The system must function effectively and reach the entire population. Furthermore, the system must address the needs of all members of the community and should be reliable.

(a) Effectiveness

While designing an early warning system, following questions should carefully be considered as to ensure effectiveness of the system:

- Does the system ensure that the early warning messages reach the last and most vulnerable person of the community?
- Does the early warning message help reduce disaster risks?
- Is early warning message beneficial for saving human, physical and financial capital of the community?

- Is the system well managed and are the resources used in the most appropriate way?

(b) Efficiency

Any established Community Based Early Warning Systems (CBEWS) should function properly which means it should be managed efficiently and be effective in protecting life and property during the time of the flood. Efficiency of any CBEWS can be assessed by following:

- Are there prompt and effective decision-making policies and systems in place to achieve its objectives?
- Will people have a positive perception on the immediate danger and is the level of understanding about hazard type appropriate?
- Can the EW messages be issued on time?
- Are the early warning facilities appropriate and are decisions taken by those at risk timely?

(c) Equity

The CBEWS must address the need of all community members. Special needs of the most vulnerable groups of the community (women, people with disabilities, elderly people, and children) should be considered. The Equity of any CBEWS can be assessed by following:

- Does the system address human justice?
- Are the voices of the most vulnerable people in the community heard?
- Are the special needs of the women, elderly people, people with disabilities and children considered and addressed?

(d) Legitimacy

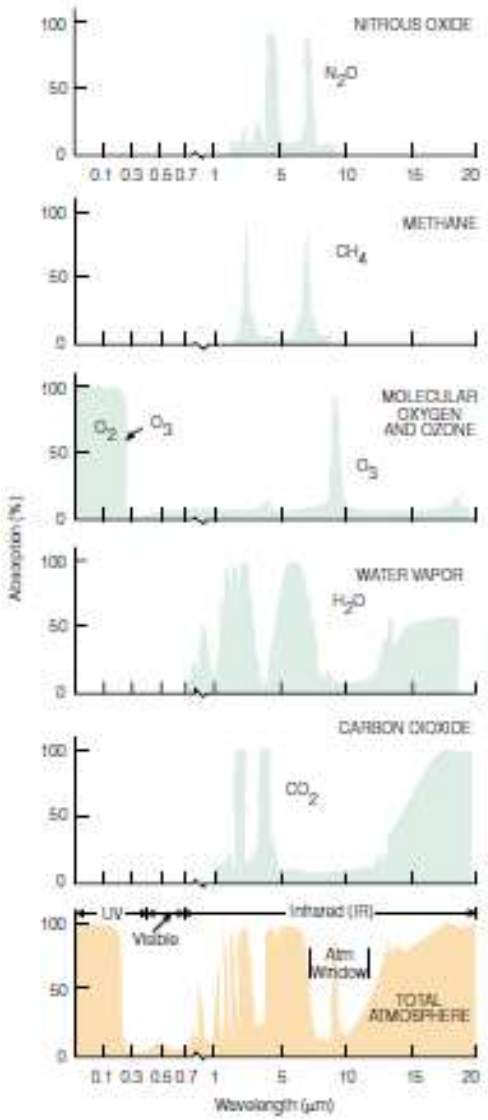
The community people take the early warning message authentic and interpret properly to cope with the situation and make responses. The early warning message should not be considered like a Nepali Folk Story. EWS could be adopted and developed as common practice and culture of the community. Legitimacy of any CBEWS can be assessed by following:

- Are the early warning messages accepted by the community or end users?
- Are there any possibilities of issuing the wrong early warning messages?
- Does the community respond to the early warnings?
- Are the local knowledge/traditional early warning practice accepted by the systems?

SESSION 9: GREENHOUSE EFFECT & CLIMATE CHANGE

9.1 Greenhouse Effect

Fig. 8.1: Absorption of radiation by gases in the atmosphere. The shaded area represents the percent of radiation absorbed. The strongest absorbers of infrared radiation are water vapor and carbon dioxide.



There are many selective absorbers in our environment. Snow, for example, is a good absorber of infrared radiation but a poor absorber of sunlight. Objects that selectively absorb radiation usually selectively emit radiation at the same wavelength. Snow is therefore a good emitter of infrared energy. At night, a snow surface usually emits much more infrared energy than it absorbs from its surroundings. This large loss of infrared radiation (coupled with the insulating qualities of snow) causes the air above a snow surface on a clear, winter night to become extremely cold. Fig. 8.1 shows some of the most important selectively absorbing gases in our atmosphere (the shaded area represents the percent of radiation absorbed by each gas at various wavelengths). Notice that both water vapor (H_2O) and carbon dioxide (CO_2) are strong absorbers of infrared radiation and poor absorbers of visible solar radiation.

Other, less important, selective absorbers include nitrous oxide (N_2O), methane (CH_4), and ozone (O3), which is most abundant in the stratosphere. As these gases absorb infrared radiation emitted from the earth's surface, they gain kinetic energy (energy of motion). The gas molecules share this energy by colliding with neighboring air molecules, such as oxygen and nitrogen (both of which are poor absorbers of infrared energy). These collisions increase the average kinetic energy of the air, which results in an increase in air temperature. Thus, most of the infrared energy emitted from the earth's surface keeps the lower atmosphere warm.

Besides being selective absorbers, water vapor and CO_2 selectively emit radiation at infrared wavelengths. This radiation travels away from these gases in all directions. A portion of this energy is radiated toward the earth's surface and absorbed, thus heating the ground. The earth, in turn, radiates infrared energy upward, where it is absorbed and warms the lower atmosphere. In this way, water vapor and CO_2 absorb and radiate infrared energy and act as an insulating layer around the earth, keeping part of the earth's infrared radiation from escaping rapidly into space. Consequently, the earth's surface and the lower atmosphere are much warmer than they would be if these selectively absorbing gases were not present. In fact, as we saw earlier, the earth's mean radiative equilibrium temperature without CO_2 and water vapor would be around −18°C (0°F), or about 33°C (59°F) lower than at present.

The absorption characteristics of water vapor, CO2, and other gases such as methane and nitrous oxide (Fig. 8.1), were, at one time, thought to be similar to the glass of a florist's greenhouse. In a greenhouse, the glass allows visible radiation to come in, but inhibits to some degree the passage of outgoing infrared radiation. For this reason, the behavior of the water vapor and CO2 in the atmosphere is popularly called the greenhouse effect.

However, studies have shown that the warm air inside a greenhouse is probably caused more by the air's inability to circulate and mix with the cooler outside air, rather than by the entrapment of infrared energy. Because of these findings, some scientists insist that the greenhouse effect should be called the atmosphere effect. To accommodate everyone, we will usually use the term atmospheric greenhouse effect when describing the role that water vapor and CO2 play in keeping the earth's mean surface temperature higher than it otherwise would be.

Looking again at Fig. 8.1 and observe that, in the bottom diagram, there is a region between about 8 and 11 μm where neither water vapor nor CO2 readily absorb infrared radiation. Because these wavelengths of emitted energy pass upward through the atmosphere and out into space, the wavelength range (between 8 and 11 μm) is known as the atmospheric window. At night, clouds can enhance the atmospheric greenhouse effect. Tiny liquid cloud droplets are selective absorbers in that they are good absorbers of infrared radiation but poor absorbers of visible solar radiation. Clouds even absorb the wavelengths between 8 and 11 μm, which are otherwise "passed up" by water vapor and CO2. Thus, they have the effect of enhancing the atmospheric greenhouse effect by closing the atmospheric window.

Clouds are also excellent emitters of infrared radiation. Their tops radiate infrared energy upward and their bases radiate energy back to the earth's surface where it is absorbed and, in a sense, reradiated back to the clouds. This process keeps calm, cloudy nights warmer than calm, clear ones. If the clouds remain into the next day, they prevent much of the sunlight from reaching the ground by reflecting it back to space. Since the ground does not heat up as much as it would in full sunshine, cloudy, calm days are normally cooler than clear, calm days. Hence, the presence of clouds tends to keep nighttime temperatures higher and daytime temperatures lower.

In summary, the atmospheric greenhouse effect occurs because water vapor, CO2, and other trace gases are selective absorbers. They allow most of the sun's radiation to reach the surface, but they absorb a good portion of the earth's outgoing infrared radiation, preventing

it from escaping into space (Fig. 8.2). It is the atmospheric greenhouse effect, then, that keeps the temperature of our planet at a level where life can survive. The greenhouse effect is not just a "good thing"; it is essential to life on earth.

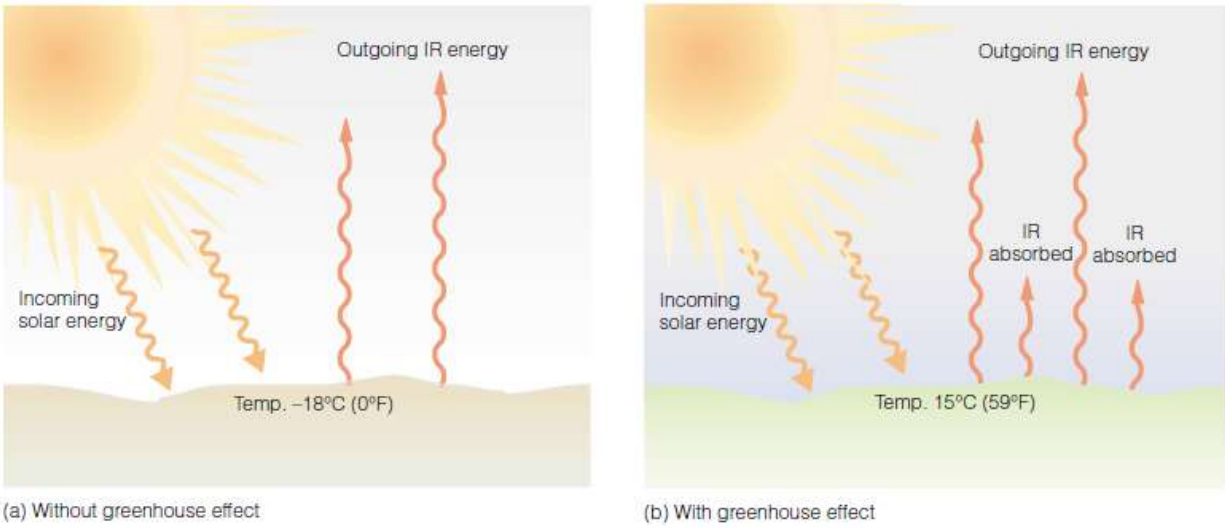


Fig. 8.2: Sunlight warms the earth's surface only during the day, whereas the surface constantly emits infrared radiation upward during the day and at night. (a) Near the surface without water vapor, CO_2, and other greenhouse gases, the earth's surface would constantly emit infrared radiation (IR) energy; incoming energy from the sun would be equal to outgoing IR energy from the earth's surface. Since the earth would receive no IR energy from its lower atmosphere (no atmospheric greenhouse effect), the earth's average surface temperature would be a frigid -18°C (0°F). (b) With greenhouse gases, the earth's surface receives energy from the sun and infrared energy from its atmosphere. Incoming energy still equals outgoing energy, but the added IR energy from the greenhouse gases raises the earth's average surface temperature about 33°C, to a comfortable 15°C (59°F).

9.2 Enhancement of Greenhouse Effect

In spite of the inaccuracies that plague temperature measurements, studies suggest that for the past 100 years or so, the earth's surface air temperature has been undergoing a slight warming of about 0.6°C (about 1°F). There are scientific computer models, called general circulation models (GCMs) that mathematically simulate the physical processes of the atmosphere and oceans. These models (also referred to as climate models) predict that if such a warming should continue unabated, we would be irrevocably committed to some measure of climate change, notably a shift of the world's wind patterns that steer the rain-producing storms across the globe.

Many scientists believe that the main cause of this global warming is the greenhouse gas CO2, whose concentration has been increasing primarily due to the burning of fossil fuels and deforestation. However, in recent years, increasing concentration of other greenhouse gases, such as methane (CH4), nitrous oxide (N2O), and chlorofluorocarbons (CFCs), has collectively been shown to have an effect almost equal to CO2.

Fig. 8.1 noticed that both CH4 and N2O absorb strongly at infrared wavelengths. Moreover, a particular CFC (CFC-12) absorbs in the region of the atmospheric window between 8 and 11 μm. Thus, in terms of its absorption impact on infrared radiation, the addition of a single CFC-

12 molecule to the atmosphere is equivalent to adding 10,000 molecules of CO2. Overall, water vapor accounts for about 60% of the atmospheric greenhouse effect, CO2 accounts for about 26%, and the remaining greenhouse gases contribute about 14%.

Presently, the concentration of CO2 in a volume of air near the surface is about 0.037 percent. Climate models predict that doubling this amount could cause the earth's average surface temperature to rise on average 2.5 degrees Celsius by the end of the twenty-first century. How can doubling such a small quantity of CO2 and adding miniscule amounts of other greenhouse gases bring about such a large temperature increase?

Mathematical climate models predict that rising ocean temperatures will cause an increase in evaporation rates. The added water vapor- the primary greenhouse gas- will enhance the atmospheric greenhouse effect and double the temperature rise, in what is known as a positive feedback. But there are other feedbacks to consider. The two potentially largest and least understood feedbacks in the climate system are the clouds and the oceans. Clouds can change area, depth, and radiation properties simultaneously with climatic changes. The net effect of all these changes is not totally clear at this time. Oceans, on the other hand, cover 70 percent of the planet. The response of ocean circulations, ocean temperatures, and sea ice to global warming will determine the global pattern and speed of climate change. Unfortunately, it is not now known how quickly or in what direction each of these will respond.

Satellite data from the Earth Radiation Budget Experiment (ERBE) suggest that clouds overall appear to cool the earth's climate, as they reflect and radiate away more energy than they retain. (The earth would be warmer if clouds were not present.) So, an increase in global cloudiness (if it were to occur) might offset some of the global warming brought on by an enhanced atmospheric greenhouse effect. Therefore, if clouds were to act on the climate system in this manner, they would provide a negative feedback on climate change.

Uncertainties unquestionably exist about the impact that increasing levels of CO2 and other trace gases will have on enhancing the atmospheric greenhouse effect. Nonetheless, many (but not all) scientific studies suggest that increasing the concentration of these gases in our atmosphere will lead to global-scale climatic change by the end of the twenty-first century. Such change could adversely affect water resources and agricultural productivity.

9.3 The Earth's Changing Climate

The earth's climate is always changing. 18,000 years ago, the earth was in the grip of a cold spell, with alpine glaciers extending their ice fingers down river valleys and huge ice sheets (continental glaciers) covering vast areas of North America and Europe. The ice measured several kilometers thick and extended as far south as New York and the Ohio River Valley. Perhaps the glaciers advanced 10 times during the last 2 million years, only to retreat. In the warmer periods, between glacier advances, average global temperatures were slightly higher than at present. Hence, some scientists feel that we are still in an ice age, but in the comparatively warmer part of it. Presently, glaciers cover only about 10 percent of the earth's land surface. Most of this ice is in the Greenland and Antarctic ice sheets. If global

temperatures were to rise enough so that all of this ice melted, the level of the ocean would rise about 65 m (213 ft).

Many major cities (such as New York, Tokyo, and London) would be inundated. Even a rise in global temperature of several degrees Celsius might be enough to raise sea level by about half a meter or so, flooding coastal lowlands.

9.4 Climate Through Ages

Throughout much of the earth's history, long before humanity came onto the scene, the global climate was much warmer than now, with the global mean temperature perhaps between 8°C and 15°C warmer than it is today. During most of this time, the polar regions were free of ice. These comparatively warm conditions, however, were interrupted by several periods of glaciation. Geologic evidence suggests that one glacial period occurred about 700 million years ago (m.y.a.) and another about 300 m.y.a. The most recent one—the Pleistocene epoch or, simply, the Ice Age- began about 2 m.y.a. Let's summarize the climatic conditions that led up to the Pleistocene. About 65 m.y.a., the earth was warmer than it is now; polar ice caps did not exist. Beginning about 55 m.y.a., the earth entered a long cooling trend. After millions of years, polar ice appeared. As average temperatures continued to lower, the ice grew thicker, and by about 10 m.y.a. a deep blanket of ice covered the Antarctic. Meanwhile, snow and ice began to accumulate in high mountain valleys of the Northern Hemisphere, and alpine, or valley, glaciers soon appeared.

About 2 m.y.a., continental glaciers appeared in the Northern Hemisphere, marking the beginning of the Pleistocene epoch. The Pleistocene, however, was not a period of continuous glaciation but a time when glaciers alternately advanced and retreated (melted back) over large portions of North America and Europe. Between the glacial advances were warmer periods called interglacial periods, which lasted for 10,000 years or more. The most recent North American glaciers reached their maximum thickness and extent about 18,000–22,000 years ago (y.a.). At that time, the sea level was perhaps 120 m (395 ft) lower than it is now. The lower sea level exposed vast areas of land, such as the Bering land bridge (a strip of land that connected Siberia to Alaska), which allowed human and animal migration from Asia to North America.

The ice began to retreat about 14,000 y.a. as surface temperatures slowly rose. Then, about 11,000 y.a., the average temperature suddenly dropped, and northeastern North America and northern Europe reverted back to glacial conditions. About 1000 years later, the cold spell (known as the Younger-Dryas) ended abruptly, and by 8000 y.a. the continental ice sheets over North America were gone. From about 6000–5000 y.a., the climate was probably 1°C warmer than at present. This time frame represents the warmest of the current interglacial period, or Holocene epoch. For this reason, the warm spell is referred to as the mid-Holocene maximum and, because this warm period favored the development of plants, it is also known as the climatic optimum. About 5000 y.a., a cooling trend set in, during which extensive alpine glaciers returned, but not continental ice sheets.

It is interesting to note that ice core data from Greenland reveal that rapid shifts in climate (from ice age conditions to a much warmer state) took place in as little as three years over central Greenland around the end of the Younger-Dryas. The data also reveal that similar rapid shifts in climate occurred several times toward the end of the Ice Age.

9.5 Climate During the last 1000 Years

About 1000 y.a., the Northern Hemisphere was relatively warm and dry. During this time, vineyards flourished, and wine was produced in England, indicating warm, dry summers and the absence of cold springs. It was during this warm, tranquil period of several hundred years (known as the medieval climatic optimum*) that the Vikings colonized Iceland and Greenland. Sometime around A.D. 1200, the mild climate of western Europe began to show extreme variations. For several hundred years the climate grew stormy. Both great floods and great droughts occurred. Extremely cold winters were followed by relatively warm ones.

During the cold spells, the English vineyards and the Viking settlements suffered. Europe experienced several famines during the 1300s. Around 1400 to 1550, the climate moderated. However, starting in the middle 1550s, the average temperature began to drop. This cooling trend (which continued for almost 300 years) is known as the Little Ice Age. During this time, the global mean temperature dropped by about 0.5°C, which allowed alpine glaciers to increase in size and advance down river canyons. Winters were long and severe, summers short and wet. The vineyards in England vanished and farming became impossible in the more northerly latitudes. Cut off from the rest of the world by an advancing ice pack, the Viking colony in Greenland perished.

During the Little Ice Age, one particular year stands out: 1816. In Europe that year, bad weather contributed to a poor wheat crop, and famine spread across the land. In North America, unusual blasts of cold arctic air moved through Canada and the northeastern United States between May and September. The cold spells brought heavy snow in June and killing frosts in July and August. In the warmer days that followed each cold snap, farmers replanted, only to have another cold outbreak damage the planting. The year 1816 has come to be known as "the year without a summer" or "eighteen hundred and froze-to-death." The unusually cold summer was followed by a bitterly cold winter. In the late 1800s, the average global temperature began to rise (see Fig. 14.5). From about 1900 until about 1940, the average temperature of the lower atmosphere rose nearly 0.5°C. Following the warmer period, the earth began to cool slightly over the next 25 years or so. In the late 1960s and 1970s, the cooling trend ended over most of the Northern Hemisphere. During the 1970s and into the 1980s, the average yearly temperature showed considerable fluctuation from year to year and from region to region, with the overall trend pointing to warming. The warming trend continued into the 1990s, with recent years being among the warmest of the twentieth century.

SESSION 10: DATA SIMULATION

10.1 Weather Data Simulation

Meteorology was one of the first disciplines to harness the power of computers, but the idea of using equations to predict the weather predates the computer era. It was first proposed in 1922 by the English mathematician Lewis Fry Richardson. Not having any computing power at his disposal, he estimated that making a useful, timely forecast would require 64,000 people to perform the calculations. Not a very feasible at the time, but his theory formed the basis for weather forecasting.

10.2 Numerical Weather Prediction

The forecast starts with a creation of a three-dimensional grid consisting of many data points representing the current atmospheric conditions over a region of interest, extending from the surface to the upper atmosphere. Each of these data points contains a set of atmospheric variables, e.g. temperature, pressure, wind speed and direction, humidity and so on, coming from the observational data. The interaction and behaviour of these atmospheric variables is captured by a set of equations.

These equations can be divided into two categories - dynamical and physical. The dynamical equations treat the Earth as a rotating sphere and the atmosphere as a fluid, so describing the evolution of the atmospheric flow means solving the equations of motion for a fluid on a rotating sphere. However, this is not enough to capture the complex behaviour of the atmosphere, so a number of physical equations are added to represent other atmospheric processes, such as warming, cooling, drying and moistening of the atmosphere, cloud formation, precipitation and so on.

Now, we already know that computers work in steps, so to predict a new weather state some time into the future, these equations need to be solved a number of times. The number of time steps and their length depends on a forecast timescale and type - short, medium or long term.

10.3 The Butterfly Effect

The atmosphere is a chaotic system, which means it is very susceptible to variations in the initial conditions. A tiny difference in the initial state of the atmosphere at the beginning of the simulation may lead to very different weather forecasts several days later. This concept of small causes having large effects is referred to as the butterfly effect.

We are familiar both with the term and the associated metaphor (a butterfly flapping its wings influencing a distant hurricane several weeks later). After all, it has been used not only in science but also in popular culture. The term was actually coined by Edward Lorenz, one of the pioneers of chaos theory, who encountered the effect while studying weather modelling.

In 1961 he showed that running a weather simulation, stopping it and then restarting it, produced a different weather forecast than a simulation run without stopping.

This behaviour can be explained by the way computers work - stopping of the simulation meant that the values of all variables had to be stored in a machine's memory. The problem was that the level of precision of those stored numbers was less than the precision the computer used to compute them - the numbers were being rounded. Assuming that such small differences could have no significant effect, Lorenz rounded the numbers accurate to six decimal places (e.g. 6.174122) to three decimal places (e.g. 6.174) before printing. The simulation was restarted with those slightly different numbers and the initially small differences were amplified into different weather forecasts.

Typically, to lessen the uncertainty in weather predictions, ensemble forecasting is used. In simple terms, a number of simulations are run with slightly different initial conditions and the results are combined into probabilistic forecasts, showing how likely particular weather conditions are. If results of the ensemble runs are similar, then the uncertainty is small, and if they are significantly different then the uncertainty is bigger.

10.4 Computer Simulations

Technological advances have taken us beyond the days when even an accurate one- or two-day forecast was an endeavor that required multiple teams of meteorologists and *hours* of analysis (and bitter arguments). In particular, the advent of satellites and computer models has really helped forecasters "up their game". If you watch any kind of weather coverage on television, I'd bet at some point you probably heard the on-air meteorologists say something like, "We just don't know the path of the storm at the moment because *the computer models* are not in agreement". So, let's take a closer look at what these "computer models" are, and how they work.

The formal term given to the creation of weather forecasts using a computer is numerical weather prediction (NWP). The list of instructions and calculations for creating a virtual weather forecast on a computer is the computer model (or just "model" for short). A computer knows nothing about the "weather. Computer models don't really analyze weather maps like a human meteorologist does. Instead, the computer starts with the current state of the atmosphere (using weather observations from surface weather stations, weather balloons, satellites, etc.) and uses mathematical equations that describe horizontal and vertical air motions, temperature changes, moisture processes, etc. to calculate what the atmosphere might look like at some future time.

The language of the atmosphere is mathematics, and the behavior of atmosphere depends on radiation budgets, advection, wind speed, surface pressure, relative humidity, and divergence/convergence can all be described by equations. Calculus and differential equations can better analyze and "speak the language" of the atmosphere. Some of the equations in a computer model are pretty straightforward and all of the inputs to the equations are easily measured. Other atmospheric processes are so complex, however, that

their equations must be simplified in the model. Regardless of their complexity, these equations can be used together to predict the future state of the atmosphere.

Computer models have to incorporate hundreds of complicated equations in *three* dimensions in order to predict the weather. Therefore, not surprisingly, it takes a lot of computing power to run a computer model that can predict the future state of the atmosphere. Indeed, the array of supercomputers at the National Weather Service that runs United States government's suite of computer models can do quadrillions of calculations per second.

Fig. 10.1: The National Weather Service's array of supercomputers that run numerical weather prediction models can do quadrillions of calculations per second

Computer models create weather forecasts over their designated domain, or area of the Earth that they cover. Some computer models produce highly-detailed forecasts over very small domain, but most computer models create forecasts over larger domains, perhaps for North America and surrounding waters, or even over the entire globe (models with global domains are usually referred to as "global models").

When a model is run, it starts with its set of initialization data, which is the computer's representation of the state of the atmosphere at the time the computer model run begins. The details of model initialization are pretty complicated, but it involves a mathematical scheme to synthesize surface and upper-air observations, satellite data, and other model data (to fill in the gaps where no observations exist) to create a complete representation of the initial state of the atmosphere at the time the model is run. The initialization process is far from perfect, and we'll explore some consequences of that later.

Using the initialization data and the equations the model is programed with to simulate atmospheric processes, the model calculates values for temperature, dew point, wind speed and direction, vertical air motions, etc. at some short time in the future (say, a couple of minutes) at many points in its domain. The model then takes those predictions and calculates values for the next forecast time (perhaps a couple more minutes into the future), and on and on, until the model run ends. Models making very detailed predictions over small domains may

only make forecasts a day or so into the future, while larger domain models, like global models, may make predictions ten days or more into the future. Even with the computers performing quadrillions of calculations per second, a global weather model usually takes a few hours to complete its run.

Computer models that predict the future state of the atmosphere are complex. But while numerical weather prediction models can create realistic predictions of the future state of the atmosphere, the predictions are imperfect. In fact, computer model forecasts can be very wrong, especially further into the future. There's no doubt that computers have helped revolutionize weather forecasting, making weather forecasts for anywhere in the world easily accessible and reasonably accurate several days into the future (most of the time). But, for all of the wonders of computer models, their forecasts always contain errors. British statistician George Box said, "All models are wrong, but some are useful." A number of sources of error are present in all computer model forecasts.

10.4.1 Initialization Errors

In order to create a perfect short-term weather forecast, a computer model would need a perfect representation of the initial state of the atmosphere.

In this process precise and accurate measurement of all relevant atmospheric variables like temperature, dew point, pressure, wind speed and direction, precipitation-rate, etc. are needed continuously at every single point in the atmosphere. Such perfect and continuous measurements aren't going to happen in the foreseeable future. So, we can't perfectly measure the atmosphere everywhere all the time. The observations we *do* have aren't perfect. Occasionally instruments that measure things like pressure, temperature, and wind speed go away and take erroneous measurements. This instrumental error is unavoidable, and while meteorologists try to identify and weed out the bad data, catching all of it is practically impossible. So, a little "observational error" sneaks its way into the model's initialization.

Another type of observational error arises from the way observational data are spatially distributed. It turns out that to best solve the equations that predict the future state of the atmosphere and data need to be organized in some evenly spaced manner. Furthermore, the more closely spaced the data is the better.

10.4.2. Computational Errors

Errors get introduced into computer-model forecasts within the calculations themselves. For starters, rounding numbers can introduce error. Also, each calculation is performed for some "time step" into the future (maybe a minute or two), and then the values calculated for that time are used as the basis for calculations another virtual minute or two into the future. But, skipping calculations for the times in between each "time step" also introduces errors. These errors could be reduced by making the model's time step shorter (leaving less time between the calculations), but that requires more computing power. Furthermore, computer models perform their calculations only at certain spots on the virtual Earth, and then interpolate the results to cover the whole globe, further introducing error. Again, this error could be reduced

by performing the calculations at more sites that are closer together but doing so requires more computing speed and power.

10.4.3 Oversimplifications (Parameterizations)

Some processes are oversimplified in the model because of their immense complexity. The details of radiation budgets and turbulent air motions near the ground, along with atmospheric convection (in some models) are so complex that we cannot model them accurately. Just think of the complexity of the surface of the Earth, for example, with parking lots, agricultural fields, forests, buildings, mountains, etc. Each nuance affects the local radiation budget and the way air moves over the ground (and humans can change the landscape with urban development). So, very small-scale atmospheric processes like radiation budgets near the ground tend to be "parameterized" in the models. Formally, a parameterization is an oversimplified way to simulate a process. Since parameterizations imperfectly simulate atmospheric processes, they introduce more errors into the model forecast, which tend to grow in time.

After hearing about how erroneous model forecasts can be, it might be tempting to think that model forecasts are useless. But, even though they're wrong, experienced weather forecasters who know the strengths and limitations of computer models find them very useful in making weather forecasts. Given their imperfect nature (flawed initializations and oversimplifications) meteorologists don't just run one computer model. Instead, they run many computer models! Using an "ensemble" of computer models helps meteorologists better understand the range of possibilities in the forecast, and we'll cover these so-called "ensemble forecasts" next.

10.5 Ensemble Forecasting

All weather forecasts have some uncertainty associated with them, and a chief way that weather forecasters deal with uncertainty is through the use of ensemble forecasts. At the most basic level, ensemble forecasts are just a set of different computer model forecasts all valid at the same time. Meteorologists use two basic types of ensembles:

- ensembles based on many different computer models (each with their own unique flaws)
- ensembles based on many runs of one computer model started with slightly different initialization data

The ensemble forecasts that meteorologists use fall into one of these two categories, or a combination of the two (some ensembles actually make use of different models and slightly different starting conditions). Having access to ensemble forecasts from computer models essentially gives forecasters many looks at different possibilities for an upcoming weather situation. For a pending forecast fraught with uncertainty, utilizing ensembles gives forecasters a better chance of hitting at least something (akin to minimizing forecast error). Ensemble forecasts can be used to show meteorologists the probability of various forecast events happening, so they're very useful tools.

Weather forecasters also sometimes find the ensemble mean forecast useful for evaluating the overall weather pattern one to two weeks into the future (more useful than any single model run, anyway). So, ensemble forecasts are a critical tool in modern weather forecasting but running the same model many times with slightly different initial conditions requires a lot of computing power.

10.6 Assessing Forecast Accuracy

The common ways that forecasters calculate forecast accuracy are:

Absolute error: Absolute error tells us about the forecast error based on the difference between the forecast conditions and what actually happens. Absolute error does not tell us whether the forecast was warmer or cooler than what actually happened; it only tells us about the size of the error.

Skill compared to climatology: Forecasts that have "skill" are more accurate than a generic "climatology" forecast of 30-year normal conditions. If the forecast is less accurate than using a forecast of climatological normals, then the forecast has no skill and is essentially useless.

10.7 When Forecasts Go Wrong?

- Most short-term weather forecasts (a few days or less into the future), while not perfect, are accurate enough to be useful.
- weather forecasts several days into the future or more can still be accurate and useful if they're less specific about details (exact temperatures, exact timing of precipitation, exact precipitation amounts, etc.)
- the perception of forecast accuracy suffers sometimes because of the sources people use for their forecasts (apps that only show a single weather icon with no further context or that give highly specific information too far into the future, for example)