and may develop into *hurricanes*, or *typhoons*, with winds exceeding 75 mi/h (33 m/s) over areas as large as 200 mi (300 km) in diameter. *Extratropical cyclones* usually form along the boundaries between warm and cold air masses. Such cyclones are usually larger than tropical cyclones and may produce precipitation over thousands of square miles. Figures 2-10 and 2-11 show average tracks of cyclones, or storm tracks, in January and July, respectively, in the Northern Hemisphere. An *anticyclone* is an area of relatively high pressure in which the winds tend to blow spirally outward in a clockwise direction in the Northern Hemisphere. Details on the general circulation and on the structure of cyclones and anticyclones can be found in meteorological textbooks.

## 2-10 Fronts

A *frontal surface* is the boundary between two adjacent air masses of different temperature and moisture content. Frontal “surfaces” are actually layers or zones of transition. Their thickness, however, is small with respect
to dimensions of air masses. The line of intersection of a frontal surface with the earth is called a surface front. An upper-air front is formed by the intersection of two frontal surfaces aloft and hence marks the boundary between three air masses. If the air masses are moving so that warm air displaces colder air, the front is called a warm front; conversely, it is a cold front if cold air is displacing warmer air. If the front is not moving, it is called a stationary front.

The life history of a typical extratropical cyclone is shown in Fig. 2-12. For reasons not yet completely understood but with the jet stream apparently often a factor, a wave is generated on the boundary between two air masses (Fig. 2-12B). Under conditions of dynamic stability, this wave moves along the front with little change in form and with little or no precipitation. If the wave is unstable, and particularly with a jet stream aloft, it progresses through the successive stages of Fig. 2-12. In stage C the cyclone is deepening and has a well-formed warm sector. Warm air from the warm sector is forced upward along the warm-front surface, causing widespread precipita-
FIGURE 2-12
Life cycle of a Northern Hemisphere frontal cyclone: (A) surface front between cold and warm air; (B) wave beginning to form; (C) cyclonic circulation and wave have developed; (D) faster-moving cold front is overtaking retreating warm front and reducing warm sector; (E) warm sector has been eliminated, and cyclone is dissipating. (U.S. National Weather Service.)

tion ahead of the surface front. At the same time, the advancing wedge of cold air behind the cold front is lifting the warm air and causing convective showers behind the cold front. Showers also frequently occur in the warm sector.

Cold fronts move faster than warm fronts and usually overtake them (Fig. 2-12D, E). This process is called occlusion, and the resulting surface front is called an occluded front. Cloudiness associated with an occluding system is illustrated in Fig. 2-13, which also shows the position of the jet stream in relation to the cloudiness and surface position of the occluding system. As occlusion progresses, the warm sector is displaced farther from the cyclone center (Fig. 2-12E), which is eventually deprived of the warm air necessary to maintain its energy. Cold air replaces the warm, the center fills, and the occluded front is destroyed. A new cyclone center may form, however, at the peak of the remaining warm sector. The time from initial wave development to complete occlusion is usually of the order of 3 to 4 days.
FIGURE 2-13
Idealized illustration showing relation of jet stream to occluding surface system. Note that the jet stream is over the boundary between open and closed cellular clouds and crosses the occluded front just above the junction of cold and warm fronts. (U.S. National Environmental Satellite Center.)

TEMPERATURE

2-11 Measurement of Temperature
In order to measure air temperature properly, thermometers must be placed where air circulation is relatively unobstructed, and yet they must be protected from the direct rays of the sun and from precipitation. In the United
States thermometers are placed in white, louvered, wooden instrument shelters (Fig. 2-14) through which the air can move readily. The shelter location must be typical of the area for which the measured temperatures are to be representative. Because of marked vertical temperature gradients just above the soil surface, all shelters should be about the same height above the ground for recorded temperatures to be comparable. In the United States shelters are set at about 4½ ft (1.4 m) above the ground.

There are about 6000 stations in the United States for which official temperature records are compiled. Except for a few hundred stations equipped or staffed to obtain continuous or hourly temperatures, most make a daily observation consisting of the current, maximum, and minimum temperatures. The minimum thermometer, of the alcohol-in-glass type, has an index which remains at the lowest temperature occurring since its last setting. The maximum thermometer has a constriction near the bulb which prevents the mercury from returning to the bulb as the temperature falls and thus registers the highest temperature since its last setting. The thermograph, with either a bimetallic strip or a metal tube filled with alcohol or mercury for its thermometric element, makes an autographic record on a chart. Electrical-resistance thermometers, thermocouples, gas-bulb thermometers, and other types of instruments are used for special purposes. Electrical-resistance thermometers, for example, are widely used for measuring upper-air temperatures and in dew-cell hygrometers, which are described briefly in the section on humidity.
2-12 Terminology

A knowledge of terminology and methods of computation is required in order to avoid misuse of published temperature data. The terms average, mean, and normal are arithmetic means. The first two are used interchangeably, but the normal [6], generally used as a standard of comparison, is the average value for a particular date, month, season, or year over a specific 30-y period (1941 to 1970 as of 1974). Plans call for recomputing the 30-y normals every decade, dropping off the first 10 y and adding the most recent 10 y.

The mean daily temperature may be computed by several methods [7]. The most accurate practical method is to average hourly temperatures. Acceptably accurate results can be obtained by averaging 3- or 6-h observations, although the random error for an individual day with irregular variations may be of some importance, especially for 6-h observations. In some countries, climatological observations are made at hours (usually three per day: morning, noon, and evening) selected to permit computation of acceptable daily means by a formula which gives the mean as a linear function of the observed values with constants depending on observation times, time of year, and location.

In the United States the mean daily temperature is the average of the daily maximum and minimum temperatures. This yields a value usually less than a degree above the true daily average. Once-daily temperature observations are usually made about 7 A.M. or 5 P.M. Temperatures are published as of the date of the reading even though the maximum or minimum may have occurred on the preceding day. Mean temperatures computed from evening readings tend to be slightly higher than those from midnight readings. Morning readings yield mean temperatures with a negative bias, but the difference is less than that for evening readings. The maximum effect [8] on mean temperature of arbitrary changes in observation time varies with place and season, and may exceed 3 Fahrenheit (1.6 Celsius) degrees.

The normal daily temperature is the average daily mean temperature for a given date computed for a specific 30-y period. The daily range in temperature is the difference between the highest and lowest temperatures recorded on a particular day. The mean monthly temperature is the average of the mean monthly maximum and minimum temperatures. The mean annual temperature is the average of the monthly means for the year.

The degree-day is a departure of one degree for one day in the mean daily temperature from a specified base temperature. For snowmelt computations, the number of degree-days for a day is equal to the mean daily temperature minus the base temperature, all negative differences being taken as zero. The number of degree-days in a month or other time interval is the total of the daily values. Published degree-day values are for heating and cooling purposes and are based on departures below and above 65°F (18°C).
The *lapse rate*, or vertical temperature gradient, is the rate of change of temperature with height in the free atmosphere. The mean lapse rate is a decrease of about 0.7 Celsius degrees per 100 m (3.8 Fahrenheit degrees per 1000 ft) increase in height. The greatest variations in lapse rate are found in the layer of air just above the land surface. The earth radiates heat energy to space at a relatively constant rate which is a function of its absolute temperature, in kelvins.† Incoming radiation at night is less than the outgoing, and the temperature of the earth’s surface and of the air immediately above it decreases. The surface cooling sometimes leads to an increase of temperature with altitude, or *temperature inversion*, in the surface layer. This condition usually occurs on still, clear nights because there is little turbulent mixing of air and because outgoing radiation is unhampered by clouds. Temperature inversions are also observed at higher levels when warm-air currents overrun colder air.

In the daytime there is a tendency for steep lapse rates because of the relatively high temperatures of the air near the ground. This daytime heating usually destroys a surface radiation inversion by early forenoon. As the heating continues, the lapse rate in the lower layers of the air steepens until it may reach the *dry-adiabatic lapse rate* (1 Celsius degree per 100 m, or 5.4 Fahrenheit degrees per 1000 ft), which is the rate of temperature change of unsaturated air resulting from expansion or compression as the air rises (lowering pressure) or descends (increasing pressure) without heat being added or removed.

Air having a dry-adiabatic lapse rate mixes readily, whereas a temperature inversion indicates a stable condition in which warm lighter air overlies cold denser air. Under optimum surface heating conditions, the air near the ground may be heated sufficiently for the lapse rate in the lowest layers to become *superadiabatic*, i.e., exceeding 1 Celsius degree per 100 m (5.4 Fahrenheit degrees per 1000 ft). This is an unstable condition since any parcel of air lifted dry-adiabatically remains warmer and lighter than the surrounding air and thus has a tendency to continue rising.

If a parcel of saturated air is lifted adiabatically, its temperature will decrease and its water vapor will condense, releasing latent heat of vaporization. This heat reduces the cooling rate of the ascending air. Hence, the *saturated-adiabatic lapse rate* is less than the dry-adiabatic. It varies inversely with the water-vapor content, and hence temperature, of the air. The average value for the lower layers at temperatures above freezing is roughly half the dry-adiabatic. At very low temperatures or high altitudes there is little

† In meteorology, absolute temperature is usually given in degrees Kelvin, called kelvins.
difference between the two lapse rates because of the very small amounts of water vapor available.

If the moisture in the rising air is precipitated as it is condensed, the temperature of the air will decrease at the **pseudo-adiabatic lapse rate**, which differs very little from the **saturated-adiabatic**. Actually, the process is not strictly adiabatic, as heat is carried away by the falling precipitation. A layer of saturated air having a saturated- or pseudo-adiabatic lapse rate is said to be in **neutral equilibrium**. If its lapse rate is less than the saturated- or pseudo-adiabatic, the air is stable; if greater, unstable.

### 2-14 Geographic Distribution of Temperature

In general, surface air temperature tends to be highest at low latitudes and to decrease poleward (Figs. 2-15 and 2-16). However, this trend is greatly distorted by the influence of land and water masses, topography, and vegetation. In the interior of large islands and continents, temperatures are higher in summer and lower in winter than on coasts at corresponding latitudes.
Temperatures at high elevations are colder than at low levels, and southern slopes have warmer temperatures than northern slopes. The average rate of decrease of surface air temperature with elevation is usually between \( \frac{1}{2} \) and 1 Celsius degree per 100 m (3 and 5 Fahrenheit degrees per 1000 ft). Forested areas have higher daily minimum and lower daily maximum temperatures than barren areas. The mean temperature in a forested area may be 1 to 2 Celsius (2 to 4 Fahrenheit) degrees lower than in comparable open country, the difference being greater in summer.

The heat from a large city, which may roughly equal one-third of the solar radiation reaching it, produces local distortions in the temperature pattern so that temperatures recorded in cities may not represent the surrounding region. The mean annual temperature of cities averages about 1 Celsius degree (2 Fahrenheit degrees) higher than that of the surrounding region, most of the difference resulting from higher daily minima in the cities. Any comparison of city and country temperatures must allow for differences in exposure of thermometers. In cities the instrument shelters are sometimes located on roofs. On still, clear nights, when radiational cooling is particularly
effective, the temperature on the ground may be as much as 8 Celsius (15 Fahrenheit) degrees colder than that at an elevation of 30 m (100 ft). A slight gradient in the opposite direction is observed on windy or cloudy nights. Daytime maxima tend to be lower at rooftop level than at the ground. In general, the average temperature from roof exposures is slightly lower than that on the ground.

2-15 Time Variations of Temperature

In continental regions the warmest and coldest points of the annual temperature cycle lag behind the solstices by about 1 month. In the United States, January is usually the coldest month and July the warmest. At oceanic stations the lag is nearer 2 months, and the temperature difference between the warmest and coldest months is much less.

The daily variation of temperature lags slightly behind the daily variation of solar radiation. The temperature begins to rise shortly after sunrise, reaches a peak 1 to 3 h (about ½ h at oceanic stations) after the sun has reached its highest altitude, and falls through the night to a minimum about sunrise. The daily range of temperature is affected by the state of the sky. On cloudy days the maximum temperature is lower because of reduced insolation, and the minimum is higher because of reduced outgoing radiation. The daily range is also smaller over oceans.

HUMIDITY

2-16 Properties of Water Vapor

The process by which liquid water is converted into vapor is called vaporization or evaporation. Molecules of water having sufficient kinetic energy to overcome the attractive forces tending to hold them within the body of liquid water are projected through the water surface. Since kinetic energy increases and surface tension decreases as temperature rises, the rate of evaporation increases with temperature. Most atmospheric vapor is the product of evaporation from water surfaces. Molecules may leave a snow or ice surface in the same manner as they leave a liquid. The process whereby a solid is transformed directly to the vapor state, and vice versa, is called sublimation.

In any mixture of gases, each gas exerts a partial pressure independent of the other gases. The partial pressure exerted by water vapor is called vapor pressure. If all the water vapor in a closed container of moist air with an initial total pressure \( p \) were removed, the final pressure \( p' \) of the dry air alone would be less than \( p \). The vapor pressure \( e \) would be the difference between the pressure of the moist air and that of the dry air, or \( p - p' \).
Practically speaking, the maximum amount of water vapor that can exist in any given space is a function of temperature and is independent of the coexistence of other gases. When the maximum amount of water vapor for a given temperature is contained in a given space, the space is said to be saturated. The more common expression “the air is saturated” is not strictly correct. The pressure exerted by the vapor in a saturated space is called the saturation vapor pressure, which, for all practical purposes, is the maximum vapor pressure possible at a given temperature (Appendix Tables B-9 and B-10).

The process by which vapor changes to the liquid or solid state is called condensation. In a space in contact with a water surface, condensation and vaporization always go on simultaneously. If the space is not saturated, the rate of vaporization will exceed the rate of condensation, resulting in a net evaporation.† If the space is saturated, the rates of vaporization and condensation balance, provided that the water and air temperatures are the same.

Since the saturation vapor pressure over ice is less than that over water at the same temperature, the introduction of ice into a space saturated with respect to liquid water at the same or higher temperature will result in condensation of vapor on the ice.

Vaporization removes heat from the liquid being vaporized, while condensation adds heat. The latent heat of vaporization is the amount of heat absorbed by a unit mass of a substance, without change in temperature, while passing from the liquid to the vapor state. The change from vapor to the liquid state releases an equivalent amount of heat.

The heat of vaporization of water $H_v$ in calories per gram varies with temperature but can be determined accurately up to $40^\circ$C ($104^\circ$F) by

$$H_v = 597.3 - 0.564T$$  \hspace{1cm} (2-1)

where $T$ is the temperature in degrees Celsius.

The latent heat of fusion for water is the amount of heat required to convert one gram of ice to liquid water at the same temperature. When one gram of liquid water at $0^\circ$C ($32^\circ$F) freezes into ice at the same temperature, the latent heat of fusion (79.7 cal/g) is liberated.

The latent heat of sublimation for water is the amount of heat required to convert one gram of ice into vapor at the same temperature without passing through the intermediate liquid state. It is equal to the sum of the latent heat of vaporization and the latent heat of fusion. At $0^\circ$C ($32^\circ$F) it is about 677 cal/g. Direct condensation of vapor into ice at the same temperature liberates an equivalent amount of heat.

† In hydrology, net evaporation is termed simply evaporation.
The specific gravity of water vapor is 0.622 times that of dry air at the same temperature and pressure. The density of water vapor $\rho_v$ in grams per cubic centimeter is

$$\rho_v = 0.622 \frac{e}{R_g T} \quad (2-2)$$

where $T$ is the absolute temperature in kelvins and $R_g$, the gas constant, equals $2.87 \times 10^3$ when the vapor pressure $e$ is in millibars.†

The density of dry air $\rho_d$ in grams per cubic centimeter is

$$\rho_d = \frac{p_d}{R_g T} \quad (2-3)$$

where $p_d$ is the pressure in millibars.

The density of moist air is equal to the mass of water vapor plus the mass of dry air in a unit volume of the mixture. If $p_a$ is the total pressure of the moist air, $p_a - e$ will be the partial pressure of the dry air alone. Adding Eqs. (2-2) and (2-3) and substituting $p_a - e$ for $p_d$ gives

$$\rho_a = \frac{p_a}{R_g T} \left(1 - 0.378 \frac{e}{p_a}\right) \quad (2-4)$$

This equation shows that moist air is lighter than dry air.

2-17 Terminology

There are many expressions used for indicating the moisture content of the atmosphere. Each serves special purposes, and only those expressions common to hydrologic uses are discussed here. Vapor pressure $e$, usually expressed in millibars but sometimes in inches of mercury, is the pressure exerted by the vapor molecules. It is most commonly used in meteorology and hydrology to denote the partial pressure of the water vapor in the atmosphere. The saturation vapor pressure $e_s$ is the maximum vapor pressure in saturated space and is a function of temperature alone. At any given temperature below the freezing point the saturation vapor pressure over liquid water is slightly greater than that over ice. The difference is at a maximum at about $-12^\circ$C ($10^\circ$F), but the ratio of the vapor pressures increases with decreasing temperature. Vapor pressure over water is generally used for most meteorological purposes regardless of temperature.

† The millibar is the standard unit of pressure in meteorology. It is equivalent to a force of 1000 dynes/cm², 0.0143 lb/in², or 0.0295 inches of mercury (abbreviated in. Hg). Mean sea-level air pressure is 1013 millibars.
The computation of saturation vapor pressure is somewhat complicated, and its values are obtained generally from psychrometric tables such as those in Appendix B and in Smithsonian Meteorological Tables [9], which are based on the Goff-Gratch formula [10]. This formula yields values of saturation vapor pressure over water that are approximated to within 1 percent in the range of $-50$ to $+55^\circ C$ ($-58$ to $+131^\circ F$) by the much simpler equation [11]

\[
es = 33.8639[(0.007387T + 0.8072)^8 - 0.000019(1.8T + 48) + 0.001316]\]

(2-5)

where $e_s$ is in millibars and $T$ is in degrees Celsius.

It is sometimes necessary to convert vapor pressure over water to that over ice, or vice versa. Ratios for effecting these conversions are found in Table B-11. The following equation [12], derived for making the conversions on a computer, yields ratios to within 0.1 percent for the range 0 to $-50^\circ C$ (32 to $-58^\circ F$):

\[
\frac{e_{s, \text{ice}}}{e_{s, \text{water}}} \approx 1 + 0.00972T + 0.000042T^2
\]

(2-6)

where $e_s$ is in millibars and $T$ is in degrees Celsius.

The *dewpoint* $T_d$ is the temperature at which space becomes saturated when air is cooled under constant pressure and with constant water-vapor content. It is the temperature having a saturation vapor pressure $e_s$ equal to the existing vapor pressure $e$.

When the relative humidity is known, the dewpoint can be approximated to within 0.3 Celsius (0.5 Fahrenheit) degrees in the temperature range of $-40$ to $50^\circ C$ ($-40$ to $122^\circ F$) by the following formula [13], which yields the dewpoint depression to be subtracted from the temperature to obtain the dewpoint

\[
T - T_d \approx (14.55 + 0.114T)X + [(2.5 + 0.007T)X]^3
\]

\[
+ (15.9 + 0.117T)X^{14}
\]

(2-7)

where $T$ is in degrees Celsius and $X$ is the complement of the relative humidity $f$ expressed as a decimal fraction, or $X = 1.00 - f/100$.

The *relative humidity* $f$ is the percentage ratio of the actual to the saturation vapor pressure and is therefore a ratio of the amount of moisture in a given space to the amount the space could contain if saturated.

\[
f = 100 \frac{e}{e_s}
\]

(2-8)
The computation of saturation vapor pressure is somewhat complicated, and its values are obtained generally from psychrometric tables such as those in Appendix B and in Smithsonian Meteorological Tables [9], which are based on the Goff-Gratch formula [10]. This formula yields values of saturation vapor pressure over water that are approximated to within 1 percent in the range of \(-50\) to \(+55^\circ C\) \((-58\) to \(+131^\circ F\)) by the much simpler equation [11]

\[
e_s \approx 33.8639[(0.007387T + 0.8072)^8 - 0.000019(1.8T + 48) + 0.001316]
\]  

(2-5)

where \(e_s\) is in millibars and \(T\) is in degrees Celsius.

It is sometimes necessary to convert vapor pressure over water to that over ice, or vice versa. Ratios for effecting these conversions are found in Table B-11. The following equation [12], derived for making the conversions on a computer, yields ratios to within 0.1 percent for the range \(0\) to \(-50^\circ C\) \((32\) to \(-58^\circ F)\):

\[
\frac{e_{s,\text{ice}}}{e_{s,\text{water}}} \approx 1 + 0.00972T + 0.000042T^2
\]  

(2-6)

where \(e_s\) is in millibars and \(T\) is in degrees Celsius.

The **dewpoint** \(T_d\) is the temperature at which space becomes saturated when air is cooled under constant pressure and with constant water-vapor content. It is the temperature having a saturation vapor pressure \(e_s\) equal to the existing vapor pressure \(e_s\).

When the relative humidity is known, the dewpoint can be approximated to within 0.3 Celsius \((0.5\) Fahrenheit) degrees in the temperature range of \(-40\) to \(50^\circ C\) \((-40\) to \(122^\circ F)\) by the following formula [13], which yields the dewpoint depression to be subtracted from the temperature to obtain the dewpoint

\[
T - T_d \approx (14.55 + 0.114T)X + [(2.5 + 0.007T)X]^3
+ (15.9 + 0.117T)X^{14}
\]  

(2-7)

where \(T\) is in degrees Celsius and \(X\) is the complement of the relative humidity \(f\) expressed as a decimal fraction, or \(X = 1.00 - f/100\).

The **relative humidity** \(f\) is the percentage ratio of the actual to the saturation vapor pressure and is therefore a ratio of the amount of moisture in a given space to the amount the space could contain if saturated.

\[
f = 100 \frac{e}{e_s}
\]  

(2-8)
Relative humidity can also be computed directly from air temperature $T$ and dewpoint $T_d$ by an approximate formula [14] in convenient form for computer use:

$$f \approx \left( \frac{112 - 0.1T + T_d}{112 + 0.9T} \right)^8$$  \hspace{1cm} (2-9)$$

where temperatures are in degrees Celsius. The formula approximates relative humidity to within 1.2 percent in the meteorological range of temperatures and humidities and to within 0.6 percent in the range of $-25$ to $45^\circ C$ ($-13$ to $113^\circ F$).

Psychrometric tables like those of Appendix B usually give dewpoint and relative humidity as a function of air temperature and wet-bulb depression, i.e., the difference between air and wet-bulb temperatures (Sec. 2-18).

The specific humidity $q_h$, usually expressed in grams per kilogram, is the mass of water vapor per unit mass of moist air:

$$q_h = 622 \frac{e}{p_a - 0.378e} \approx 622 \frac{e}{p_a}$$  \hspace{1cm} (2-10)$$

where $p_a$ is the total pressure of the air in millibars.

The mixing ratio $w_r$ is the mass of water vapor per unit mass of perfectly dry air in a humid mixture. Usually expressed in grams per kilogram of dry air, it is given by

$$w_r = 622 \frac{e}{p_a - e}$$  \hspace{1cm} (2-11)$$

The total amount of water vapor in a layer of air is often expressed as the depth of precipitable water $W_p$, in millimeters or inches, even though there is no natural process capable of precipitating the entire moisture content of the layer. The amount of precipitable water in any air column of considerable height is computed [15] by increments of pressure or height from surface and upper-air observations of temperature, humidity, and pressure. A convenient formula [16] for computing $W_p$ in millimeters is

$$W_p = \sum 0.01 \bar{q}_h \Delta p_a$$  \hspace{1cm} (2-12)$$

where the pressure $p_a$ is in millibars and $\bar{q}_h$, in grams per kilogram, is the average of the specific humidities at the top and bottom of each layer. Precipitable water may be approximated conveniently by nomograms [17] using dewpoints at specific pressure levels. Less accurate approximations, which may be considerably in error under certain conditions, can be made from surface dewpoints [18] or vapor pressure and assumed temperature and humidity lapse rates or relations based on observed precipitable water. Figure 2-17 gives the depth of precipitable water in a column of saturated air.
FIGURE 2-17
Depths of precipitable water in a column of air of any height above the 1000-millibar level as a function of the 1000-millibar dewpoint, assuming saturation and pseudo-adiabatic lapse rate. (U.S. National Weather Service.)
with its base at the 1000-millibar level and its top anywhere up to 200 milli-
bars. Tables for computing precipitable water in various layers of the
saturated atmosphere are available [19].

2-18 Measurement of Humidity

In general, measurements of humidity in the surface layers of the atmosphere
are made with a psychrometer, which consists of two thermometers, one with
its bulb covered by a jacket of clean muslin saturated with water. The
thermometers are ventilated by whirling or by use of a fan. Because of the
cooling effect of evaporation, the moistened, or wet-bulb, thermometer reads
lower than the dry, the difference in degrees being known as the wet-bulb
depression. The air and wet-bulb temperatures are used to obtain various
expressions of humidity by reference to psychrometric tables (Appendix B).

The hair hygrometer makes use of the fact that the length of a hair
varies with relative humidity. The changes are transmitted to a pointer
indicating the relative humidity on a graduated scale. The hair hygrograph
is a hair hygrometer operating a pen marking a trace on a chart. The
hygrothermograph, combining the features of both the hair hygrograph and
thermograph, records both relative humidity and temperature on one chart.

A dewpoint hygrometer, which measures dewpoint directly and is used mostly
for laboratory purposes, consists of a highly polished metal vessel containing
a suitable liquid which is cooled by any of several methods. The temperature
of the liquid at the time condensation begins to occur on the exterior of the
metal vessel is the dewpoint. The dew-cell hygrometer measures the dewpoint
by regulating the temperature of a saturated aqueous lithium chloride
solution so that the water-vapor pressure of the solution is equal to that of
the surrounding atmosphere. The spectral hygrometer measures the selective
absorption of light in certain bands of the spectrum by water vapor. With
the sun as a light source, it has been used to measure total atmospheric
moisture [20]. Other humidity instruments have been developed for special
purposes but are not generally used in routine operational activities.

Measurement of humidity is one of the least accurate instrumental
procedures in meteorology. The standard psychrometer invites many obser-
vational errors. The two thermometers double the chance of misreading. At
low temperatures, misreading by a few tenths of a degree can lead to absurd
results. There is always the chance that the readings are not made when the
wet-bulb thermometer is at its lowest temperature. In addition, there are
efforts with a positive bias resulting from insufficient ventilation, dirty or too
thick muslin, and impure water.

Any instrument using a hair element is subject to appreciable error.
The hair expands with increasing temperature, and its response to changes in
humidity is very slow, the lag increasing with decreasing temperature until it becomes almost infinite at about $-40^\circ$C ($-40^\circ$F). This can lead to significant error in upper-air soundings, where large ranges in temperature and sharp variations of humidity with altitude are observed. Consequently, sounding instruments are now equipped with electrical hygrometers using various humidity elements. A carbon element [21] has been found most satisfactory of those tested so far.

At freezing temperatures there is some uncertainty whether the dew-point or frost point is being measured by dewpoint hygrometers. The difference may lead to appreciable errors in computing relative humidity [22] and vapor pressure.

2-19 Geographic Distribution of Humidity

Atmospheric moisture tends to decrease with increasing latitude, but relative humidity, being an inverse function of temperature, tends to increase. Atmospheric moisture is greatest over oceans and decreases with distance inland. It also decreases with elevation and is greater over vegetation than over barren soil. The distribution of the average precipitable water over the Northern Hemisphere for 1958 is shown in Fig. 2-18. The distribution depicted is fairly representative of the mean annual pattern since there is generally relatively little year-to-year variation.

2-20 Time Variations in Humidity

Like temperature, atmospheric water vapor is at a minimum in winter and at a maximum in summer. In the Northern Hemisphere the driest months are January and February, and the most humid are July and August. In the middle and high latitudes average monthly precipitable water over continental interiors in the driest months is about half the mean annual; in the most humid months, about twice the mean annual. The seasonal variation is much less over oceans and coastal areas and is at a minimum over tropical seas. Unlike actual water vapor content, relative humidity is at a minimum in summer and at a maximum in winter.

The diurnal variation of atmospheric moisture is normally small, except where land and sea breezes bring air of differing characteristics. Near the ground surface, condensation of dew at night and reevaporation during the day may result in a minimum moisture content near sunrise and a maximum by noon. Relative humidity, of course, behaves in a manner opposite to that of temperature, being at a maximum in the early morning and at a minimum in the afternoon.
FIGURE 2-18

Wind, which is air in motion, is a very influential factor in several hydro-meteorological processes. Moisture and heat are readily transferred to and from air, which tends to adopt the thermal and moisture conditions of the surfaces with which it is in contact. Stagnant air in contact with a water surface eventually assumes the vapor pressure of the surface, so that no evaporation takes place. Similarly, stagnant air over a snow or ice surface...
eventually assumes the temperature and vapor pressure of the surface, so that melting by convection and condensation ceases. Consequently, wind exerts considerable influence in evaporative and snowmelt processes. It is also important in the production of precipitation, since it is only through sustained inflow of moist air into a storm that precipitation can be maintained.

2-21 Measurement of Wind

Wind has both speed and direction. The wind direction is the direction from which it is blowing. Direction is usually expressed in terms of 16 compass points (N, NNE, NE, ENE, etc.) for surface winds, and for winds aloft in degrees from north, measured clockwise. Wind speed is usually given in miles per hour, meters per second, or knots (1 m/s = 2.237 mi/h = 1.944 kn, and 1 kn = 1.151 mi/h = 0.514 m/s).

Wind speed is measured by instruments called anemometers, of which there are several types. The three- or four-cup anemometer with a vertical axis of rotation is most commonly used for official observations. It tends to register too high a mean speed in a variable wind because the cups accelerate faster than they lose speed. Vertical currents (turbulence) tend to rotate the cups and cause overregistration of horizontal speeds. Most cup anemometers will not record speeds below 1 or 2 mi/h because of starting friction. The propeller anemometer has a horizontal axis of rotation. Pressure-tube anemometers, of which the Dines is the best known, operate on the pitot-tube principle.

While wind speed varies greatly with height above the ground, no standard anemometer level has been adopted. Differences in wind speed resulting from differences in anemometer height, which may range anywhere from 30 to several hundred feet above the ground, usually exceed the errors from instrumental deficiencies. However, approximate adjustment can be made for differences in height [Eq. (2-14), (2-15), or (2-16)].

2-22 Geographic Variation of Wind

In winter there is a tendency for surface winds to blow from the colder interior of land masses toward the warmer oceans (Sec. 2-8). Conversely, in summer the winds tend to blow from the cooler bodies of water toward the warmer land. Similarly, diurnal land and sea breezes may result from temperature contrasts between land and water.

On mountain ridges and summits wind speeds at 10 m (30 ft) or more above the ground are higher than in the free air at corresponding elevations because of the convergence of the air forced by the orographic barriers. On lee slopes and in sheltered valleys wind speeds are light. Wind direction is
greatly influenced by orientation of orographic barriers. With a weak pressure system, diurnal variation of wind direction may occur in mountain regions, the winds blowing upslope in the daytime and downslope at night.

Wind speeds are reduced and directions deflected in the lower layers of the atmosphere because of friction produced by trees, buildings, and other obstacles. These effects become negligible above about 600 m (2000 ft), and this lower layer is referred to as the friction layer. Over land the surface wind speed averages about 40 percent of that just above the friction layer, and at sea about 70 percent.

Because of its location in the general circulation, most of the conterminous United States has prevailing westerly winds. However, the winds are generally variable since most of the country is affected by migratory pressure systems, with winds circulating clockwise in the high-pressure areas and counterclockwise in low-pressure systems as they move across the country.

The variation of wind speed with height, or the wind profile, in the friction layer is usually expressed by one of two general relationships, namely, the logarithmic velocity profile or the power-law profile. In hydrology, the relationships are most often used to estimate the wind speed in the surface boundary layer, i.e., the thin layer of air between the ground surface and the anemometer level, usually about 10 m (30 ft) but often lower at special test sites or experimental stations. The common requirement is the wind speed above a snow or water surface for computations of snowmelt and evaporation.

One of the more common forms [23] of the logarithmic velocity profile for meteorological purposes is

\[
\frac{\bar{v}}{v_*} = \frac{1}{k} \ln \frac{z}{z_0} \quad z \geq z_0 \quad (2-13)
\]

<table>
<thead>
<tr>
<th>Type of surface</th>
<th>( z_0 ) cm</th>
<th>( z_0 ) in.</th>
<th>( v_* ) cm/s</th>
<th>( v_* ) ft/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>Very smooth (mud flats, ice)</td>
<td>0.001</td>
<td>0.0004</td>
<td>16</td>
<td>0.5</td>
</tr>
<tr>
<td>Lawn, grass up to 1 cm (0.4 in.) high</td>
<td>0.1</td>
<td>0.4</td>
<td>26</td>
<td>0.9</td>
</tr>
<tr>
<td>Downland, thin grass up to 10 cm (4 in.) high</td>
<td>0.7</td>
<td>0.28</td>
<td>36</td>
<td>1.2</td>
</tr>
<tr>
<td>Thick grass, up to 10 cm (4 in.) high</td>
<td>2.3</td>
<td>0.91</td>
<td>45</td>
<td>1.5</td>
</tr>
<tr>
<td>Thin grass, up to 50 cm (20 in.) high</td>
<td>5</td>
<td>2.0</td>
<td>55</td>
<td>1.8</td>
</tr>
<tr>
<td>Thick grass, up to 50 cm (20 in.) high</td>
<td>9</td>
<td>3.5</td>
<td>63</td>
<td>2.1</td>
</tr>
</tbody>
</table>

where \( \bar{v} \) is the mean wind speed for at least a few minutes at height \( z \) above the ground; \( k \) is the von Kármán constant, generally taken equal to 0.4; \( z_0 \) is the roughness length, which is a measure of surface roughness and presumably the height at which wind speed becomes zero and must therefore be less than \( z \); and \( \nu \) is the friction velocity, which is equal to \( \sqrt{\tau/\rho} \), \( \tau \) being the shear stress, or Reynolds stress, and \( \rho \) the air density.

In meteorological investigations of the surface boundary layer, \( \bar{v} \) is generally considered independent of height, with the surface value assumed to apply throughout the layer, that is, \( \bar{v} = \bar{v}_0 \). Hence, the friction velocity \( \nu \) depends on the nature of the surface and the mean wind speed \( \bar{v} \). It usually ranges [24] from about 3 to 12 percent of the mean wind speed \( \bar{v} \), the lower values being associated with smooth surfaces. A rough assumption sometimes made in meteorological studies is that \( v \) is approximately equal to \( \bar{v}/10 \). Some measurements of roughness length \( z_0 \) and friction velocity \( \nu \) for a mean wind speed \( \bar{v} \) of 5 m/s (11 mi/h) at 2 m (6.5 ft) above the ground are presented in Table 2-2. A more detailed listing of \( z_0 \) for a wide range of surface roughness is given in Table 2-3.

A value of one-thirtieth the average height of surface irregularities is often taken as a fair estimate of \( z_0 \), but the data of Table 2-3 suggest that this simple approximation could be greatly in error under certain conditions, especially in the case of brush and trees. Table 2-3 shows that \( z_0 \) varies inversely with wind speed in the case of tall grasses, which tend to flatten out as speed increases [32]. In the case of a water surface, which roughens as wind speed increases, \( z_0 \) tends to be greater for higher wind speeds. However, it appears that there are discontinuities in this relationship, and, at least in the range of 2 to 10 m/s (4 to 22 mi/h) at 10 m (30 ft) above the surface, \( z_0 \) alternately increases and decreases as wind speed increases [33].

A convenient form [34] of the logarithmic velocity profile for relating mean wind speed \( \bar{v} \) at some height \( z \) to the measured mean wind speed \( \bar{v}_1 \) at some standard height \( z_1 \) is

\[
\frac{\bar{v}}{\bar{v}_1} = \frac{\ln \left( \frac{z}{z_0} + 1 \right)}{\ln \left( \frac{z_1}{z_0} + 1 \right)}
\]  

(2-14)

Another convenient form [35] of the logarithmic velocity profile for computing mean wind speed \( \bar{v}_2 \) at some intermediate height \( z_2 \) when mean wind speeds \( \bar{v}_1 \) and \( \bar{v}_3 \) at heights \( z_1 \) and \( z_3 \) are known is

\[
\bar{v}_2 = \bar{v}_3 - (\bar{v}_3 - \bar{v}_1) \frac{\ln \left( \frac{z_3}{z_2} \right)}{\ln \left( \frac{z_3}{z_1} \right)}
\]  

(2-15)
Table 2-3 ROUGHNESS LENGTH $z_0$

<table>
<thead>
<tr>
<th>Wind speed at $z = 2$ m (6.6 ft)</th>
<th>Roughness length $z_0$</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>m/s</td>
<td>mi/h</td>
<td>cm</td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>Open water</td>
<td>2.1</td>
<td>4.7</td>
</tr>
<tr>
<td>Smooth mud flats</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Smooth snow on short grass</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Wet soil</td>
<td>1.8</td>
<td>4.0</td>
</tr>
<tr>
<td>Desert</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Snow on prairie</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Mown grass:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1.5 cm (0.6 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>3.0 cm (1.2 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>4.5 cm (1.8 in.)</td>
<td>2</td>
<td>4.5</td>
</tr>
<tr>
<td>4.5 cm (1.8 in.)</td>
<td>6-8</td>
<td>13-18</td>
</tr>
<tr>
<td>Alfalfa:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>20–30 cm (8–12 in.)</td>
<td>1.9</td>
<td>4.3</td>
</tr>
<tr>
<td>30–40 cm (12–16 in.)</td>
<td>1.9</td>
<td>4.3</td>
</tr>
<tr>
<td>Long grass:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>60–70 cm (24–28 in.)</td>
<td>1.5</td>
<td>3.4</td>
</tr>
<tr>
<td>60–70 cm (24–28 in.)</td>
<td>3.5</td>
<td>7.8</td>
</tr>
<tr>
<td>60–70 cm (24–28 in.)</td>
<td>6.2</td>
<td>13.9</td>
</tr>
<tr>
<td>Maize:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>90 cm (35 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>170 cm (67 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>300 cm (118 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Sugar cane:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>100 cm (39 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>200 cm (79 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>300 cm (118 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>400 cm (157 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Brush, 135 cm (53 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Orange orchard, 350 cm (138 in.)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Pine forest:</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5 m (16 ft)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>27 m (89 ft)</td>
<td>...</td>
<td>...</td>
</tr>
<tr>
<td>Deciduous forest, 17 m (56 ft)</td>
<td>...</td>
<td>...</td>
</tr>
</tbody>
</table>

For most meteorological purposes the power-law profile is usually expressed as

\[ \frac{\bar{v}}{\bar{v}_1} = \left( \frac{z}{z_1} \right)^k \]  

(2-16)

with exponent \( k \) varying with surface roughness and atmospheric stability and usually ranging from 0.1 to 0.6 (Table 2-4) in the surface boundary layer [36, 37].

Comparative tests of the two profiles have yielded inconclusive results. The literature suggests that the logarithmic law should be the more representative of the two relationships, but this is not supported by tests except under certain conditions. The logarithmic law has been found generally more representative of the wind profile in the lowest 5 to 8 m (15 to 25 ft) above the ground when the atmospheric temperature lapse rate was adiabatic or near adiabatic [38, 39]. However, Johnson [36] found greater wind-speed increases with height over prairie grass and a snow surface than was indicated by the logarithmic law even under adiabatic conditions.

The power-law profile is considered by some investigators [40-42] to be more representative of the wind profile in the layer from several meters to about 100 m (300 ft) above the ground. Analyzing results of his own investigations, which were based on 1-h averaging periods, and those of other investigators, DeMarrais [42] concluded that for this range of elevation

Table 2-4

<table>
<thead>
<tr>
<th>Surface</th>
<th>Height range</th>
<th>Super-adiabatic</th>
<th>Neutral</th>
<th>Stable</th>
<th>Inversion</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Meadows</td>
<td>10-70</td>
<td>33-230</td>
<td>0.25</td>
<td>0.27</td>
<td>...</td>
<td>0.61</td>
</tr>
<tr>
<td>Flat field</td>
<td>11-49</td>
<td>36-161</td>
<td>0.16</td>
<td>0.20</td>
<td>0.25</td>
<td>0.36</td>
</tr>
<tr>
<td>Grass field</td>
<td>8-120</td>
<td>26-394</td>
<td>0.14</td>
<td>0.17</td>
<td>0.27</td>
<td>0.32-0.77†</td>
</tr>
<tr>
<td>Airfield</td>
<td>9-27</td>
<td>30-89</td>
<td>0.09</td>
<td>0.08</td>
<td>0.18</td>
<td>0.22</td>
</tr>
<tr>
<td>Desert</td>
<td>6-61</td>
<td>20-200</td>
<td>0.15</td>
<td>0.18</td>
<td>0.22</td>
<td>...</td>
</tr>
<tr>
<td>Near wooded area</td>
<td>11-124</td>
<td>36-407</td>
<td>0.19</td>
<td>0.29</td>
<td>0.35</td>
<td>...</td>
</tr>
</tbody>
</table>

**Source:** Adapted by permission from [35].

† \( T_{5000} - T_{5 ft}, °F \)

<table>
<thead>
<tr>
<th>0-2</th>
<th>2-4</th>
<th>4-6</th>
<th>6-8</th>
<th>8-10</th>
<th>10-12</th>
</tr>
</thead>
<tbody>
<tr>
<td>( k )</td>
<td>0.32</td>
<td>0.44</td>
<td>0.59</td>
<td>0.62</td>
<td>0.63</td>
</tr>
</tbody>
</table>

† Captive-balloon observations; all other measurements from towers.
exponent $k$ increases with surface roughness and atmospheric stability except in the case of large superadiabatic lapse rates, when it increases with instability. Exponent $k$ also varies with height of layer considered. Higher layers have lesser values of $k$ when the lapse rate is superadiabatic, and they have greater values of $k$ when the lapse rate is adiabatic or less. For adiabatic and superadiabatic lapse rates, $k$ tends to range from 0.1 to 0.3, the variation being mainly in proportion to surface roughness. This is shown in Table 2-4, which shows also that for stable conditions $k$ tends to range from 0.2 to 0.8, the higher values being associated with temperature inversions. A value of $\frac{1}{3}$ has been found applicable for a wide range of conditions in the layer 0 to 10 m (0 to 30 ft). Over a snow surface, under conditions favoring melting, a value of $\frac{1}{3}$ may be more appropriate.

Despite differences between the two types of relationships, when $z_1/z_0$ and $k$ in the logarithmic and power-law profiles, respectively, are determined from observed winds at two levels in the surface boundary layer, say at 1 and 10 m (3 and 30 ft), differences between wind speeds indicated by the two profiles are usually within the limits of accuracy of wind-measuring devices.

2-23 Time Variation of Winds

Wind speeds are highest and most variable in winter, whereas middle and late summer is the calmest period of the year. In winter westerly winds prevail over the United States up to at least 6 km (20,000 ft), except near the Gulf of Mexico, where there is a tendency for southeasterly winds up to about 1.5 km (5000 ft). In summer, while there is still a tendency for westerly winds, there is generally more variation of direction with altitude. In the plains west of the Mississippi River there is a tendency for southerly winds up to 1.5 km (5000 ft), and on the Pacific Coast the winds at the lower altitudes are frequently from the northwest.

The diurnal variation of wind is significant only near the ground and is most pronounced during the summer. Surface-wind speed is usually at a minimum about sunrise and increases to a maximum in early afternoon. At about 300 m (1000 ft) above the ground, the maximum occurs at night and the minimum in the daytime.

REFERENCES

PROBLEMS

2-1 Show why the theoretical eastward velocity (constant angular momentum) of air at rest relative to earth's surface at the equator would be 1560 mi/h if the parcel were displaced to 60°N latitude.

How many degree-days above 32°F are there in a day with a minimum temperature of 26°F and a maximum of 48°F?

A parcel of moist air at 60°F initially at 2000 ft, mean sea level, is forced to pass over a mountain ridge at 8000 ft, mean sea level, and then descends to its original elevation. Assuming that a lift of 2000 ft produces saturation and precipitation and that the average pseudo-adiabatic lapse rate is one-half the dry-adiabatic, what is the final temperature of the parcel?

2-4 What is the heat of vaporization, in calories per gram, for water at (a) 15°C and (b) 77°C?

2-5 What is the density, in kilograms per cubic meter, of (a) dry air at 30°C and a pressure of 900 millibars and (b) moist air with relative humidity of 70 percent at the same temperature and pressure?

Assuming dry- and wet-bulb temperatures of 80 and 62°F, respectively, and using the psychrometric tables of Appendix B, determine (a) dewpoint temperature, (b) relative humidity, (c) saturation vapor pressure, and (d) actual vapor pressure, in millibars.

2-7 A radio sounding in saturated atmosphere shows temperatures of 16.0, 11.6, and 6.2°C at the 900-, 800-, and 700-millibar levels, respectively. Compute the precipitable water, in millimeters, in the layer between 900 and 700 millibars, and compare the result with that obtained from Fig. 2-17. (The temperature of 16.0°C at the 900-millibar level reduces pseudo-adiabatically to 20.0°C at 1000 millibars.)

A formula for estimating potential evapotranspiration from a lawn requires wind speed at 2 m above the surface. What is the estimated speed at this level if the roughness length is 1.0 cm and an anemometer 10 m above the ground indicates a mean wind speed of 5.0 m/s?

A pilot-balloon observation shows wind speed of 40 kn at 300 m above the ground. What is the estimated speed, in miles per hour, 30 ft above the ground indicated by the power-law profile with values of (a) 3/4 and (b) 1/2 for exponent k?
10. The saturation vapor pressure over water at 10°F is 2.40 millibars. What is it over ice at the same temperature?

11. Compute the weight, in kilograms, of 1 m³ of dry air at a temperature of (a) 0°C and a pressure of 1000 millibars and (b) 20°C at the same pressure.

12. How many calories are required to evaporate 1 gal (U.S.) of water at 70°F?
How many pounds of ice at 14°F would the same amount of heat melt?
(Specific heat of ice = 0.5.)

Anemometers mounted on a tower at 2 and 16 m above the ground indicate mean wind speeds of 2.5 and 5.0 m/s, respectively.
(a) What is the roughness length, in centimeters?
(b) Using the roughness length computed in part (a), determine the wind speed at 5 m.
(c) What does Eq. (2-15) yield for the wind speed at 5 m?

How many calories per square foot are required (a) to melt a 1-ft layer of ice with a specific gravity of 0.90 at 20°F and (b) to evaporate the resulting meltwater without raising its temperature above 32°F? (Use 0.5 for specific heat of ice.)

Anemometers at 10 and 100 m on a tower record average wind speeds of 5.0 and 10.0 m/s, respectively. Compute average speeds at 30 and 60 m using (a) Eq. (2-15) and (b) Eq. (2-16).

What is the relative humidity if air temperature and dewpoint are (a) 20 and 10°C and (b) 40 and 4°F?

What is the dewpoint for air temperature and relative humidity of (a) 15°C and 49 percent and (b) 25°F and 24 percent?

Convert Eq. (2-9) so it can be used for the Fahrenheit temperature scale.

Given a mean wind speed of 2.0 m/s at 2 m above the ground and a roughness length of 0.5 cm, compute (a) the friction velocity, in centimeters per second, and (b) the wind speed, in meters per second, at 0.5 m.