Chapter 3

ATMOSPHERIC MOISTURE

Atmospheric moisture is a key element in fire weather. It has direct effects on the flammability of forest fuels, and, by its relationship to other weather factors, it has indirect effects on other aspects of fire behavior. There is a continuous exchange of water vapor between the atmosphere and dead wildland fuels. Dry fuels absorb moisture from a humid atmosphere and give up their moisture to dry air. During very dry periods, low humidity may also affect the moisture content of green fuels. When atmospheric moisture condenses and falls as precipitation, it increases the moisture content of dead fuels, and, by replenishing soil moisture, it provides for the growth of green vegetation.

We have already seen that moisture influences all surface temperatures, including surface fuel temperatures, by controlling radiation in its vapor state and by reflecting and radiating when it is condensed into clouds. The heat energy released in condensation provides the energy for thunderstorms and the violent winds associated with them. Moisture is also necessary for the development of lightning, which in many mountainous areas is a dreaded cause of wildfire.

ATMOSPHERIC MOISTURE

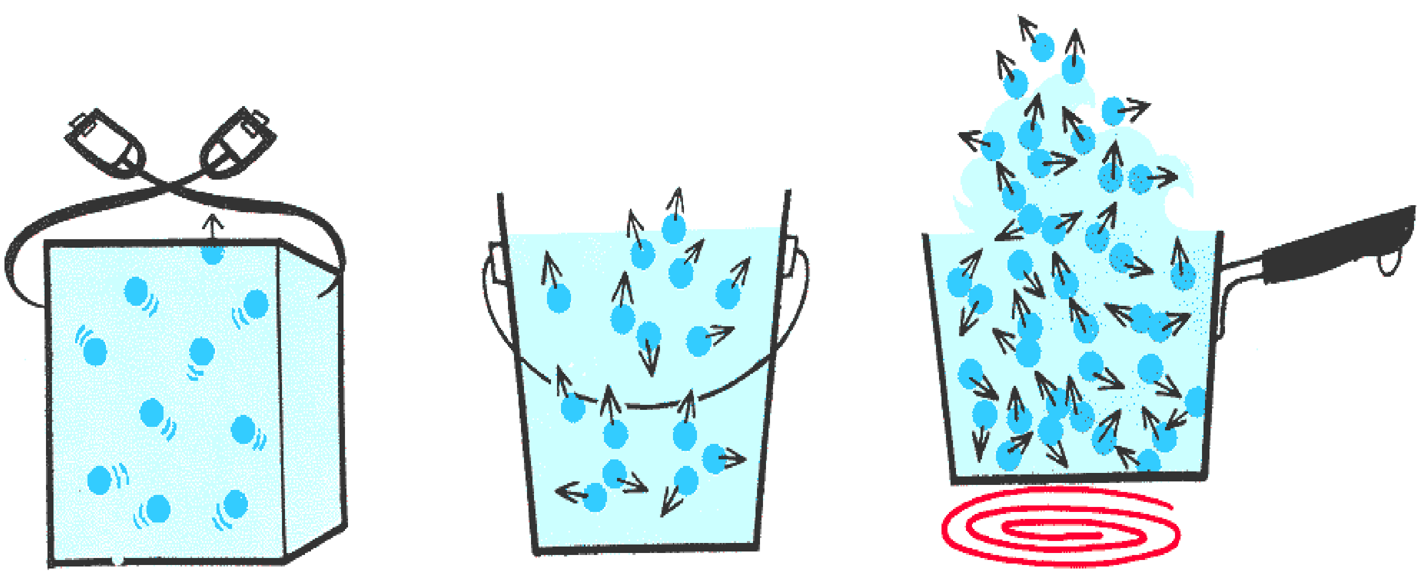
Water is always present in the lower atmosphere in one or more of its three states. It may exist as a **gas** (invisible water vapor), as a **liquid** (rain, drizzle, dew, or cloud droplets), and as a **solid** (snow, hall, sleet, frost, or ice crystals).

In its three states and in its changes from one state to another, water continually and universally influences the weather. In a later chapter we will consider atmospheric processes involving water that produce clouds and precipitation. In the present chapter we will be concerned primarily with water vapor in the atmosphere - how it gets there, how it is measured, described, and distributed, and how it varies in time and space.

WATER VAPOR IN THE ATMOSPHERE

Moisture as vapor acts the same as any other gas. It mixes with other gases in the air, and yet maintains its own identity and characteristics. It is the raw material in condensation. It stores immense quantities of energy gained in evaporation; this energy is later released in condensation. Much of the energy for thunderstorms, tornadoes, hurricanes, and other strong winds comes from the heat released when water vapor condenses. The availability of water vapor for precipitation largely determines the ability of a region to grow vegetation, which later becomes the fuel for wildland fires.

Moisture in the atmosphere is continually changing its physical state condensing into liquid, freezing into ice, melting into liquid water, evaporating into gaseous water vapor, and condensing back to liquid. These changes are all related to temperature, the gage of molecular activity in any substance. At about -460°F. (absolute zero) the molecules of all substances are motionless. As the temperature rises, they move around at increasing speeds. Water molecules move slowly at subfreezing temperatures, more rapidly at melting temperature, and still more rapidly through the boiling stage. However, at any given temperature, individual molecules, whether solid, liquid, or gas, do not have the same speeds or direction of travel. Collisions that change their speeds and directions occur continuously.

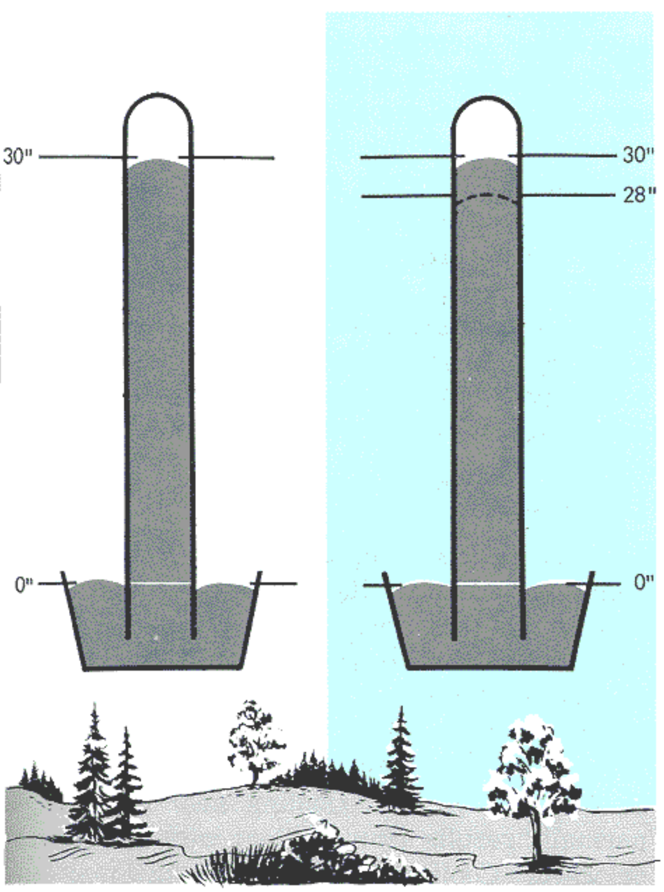


The internal pressure causing water vapor to escape from ice or liquid water varies greatly with the surface temperature; it is very small at cold temperatures and increases rapidly in liquid water through the boiling stage.

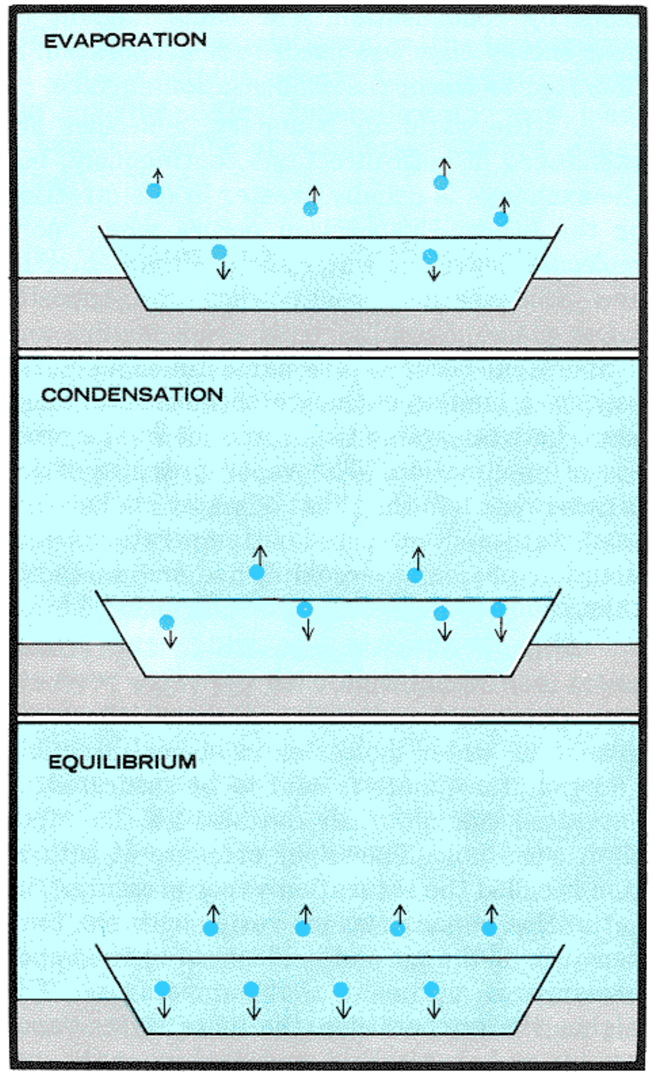
Evaporation

Some molecules momentarily acquire a very high speed from the impacts of other molecules. If this collision occurs in liquid water near the surface, and the high speed is in an outward direction, the molecules may escape into the air. This is **evaporation,** the process by which a liquid water molecule becomes a water-vapor molecule. Since molecules with the highest energy content escape, leaving behind in the liquid those with a lower energy content, the average level of energy of this liquid is decreased. The decrease in energy level results in a decrease in temperature of the liquid. Therefore, evaporation is a cooling process. Each molecule escaping into the air by a change of state takes with it nearly 1,000 times the energy needed to raise the temperature of a water molecule 1°F.

The pressure at the water-air boundary resulting from molecular motion in the direction of escape from the liquid is called the vapor pressure of water. This pressure varies only with the temperature of the water and determines



The partial pressure due to water vapor may vary from near zero in cold, dry air to about 2 inches of mercury in warm, moist air.



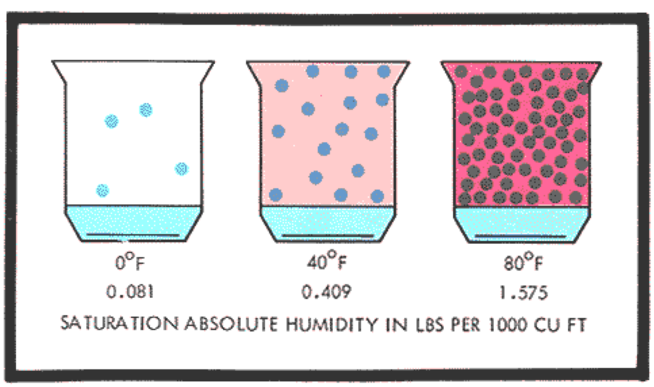
Evaporation occurs when an excess of water molecules leaves a water surface, and condensation occurs when an excess of molecular arms the liquid water. In an equilibrium condition, there is no net exchange in either direction, and the atmosphere is saturated.

the rate at which water molecules escape to the air and become vapor molecules. The water-vapor molecules, which escape to the air, displace air molecules and contribute their proportionate share to the total atmospheric pressure. This portion is called the partial pressure due to water vapor, or for simplicity, the vapor pressure.

Vapor pressure depends on the actual water vapor in the air, and it may vary from near zero in cold, dry air to about 2 inches of mercury in warm, moist air. High values can occur only in the warm, lower layers of the troposphere. The pressure produced by the vapor causes some water-vapor molecules to re-enter water surfaces by condensation. The same amount of heat energy that was needed for evaporation is liberated to warm the condensation surface.

At the water-air boundary, molecules are exchanged in both directions continuously, but the exchange is usually greater in one direction or the other. Evaporation occurs when more molecules leave the water surface than enter it, and condensation occurs when the opposite takes place. Actually, both condensation and evaporation occur at the same time. As noted earlier, a similar exchange of molecules takes place between water vapor and ice in the process of sublimation. The vapor pressure of ice is somewhat less than that of water at the same temperature. Hence, at low temperatures sublimation on ice is accomplished more readily than condensation on a water surface.

When the vapor pressure in the atmosphere is in equilibrium with the vapor pressure of a water or ice surface, there is no net exchange of water molecules in either direction, and the atmosphere is said to be saturated. A saturated volume of air contains all the vapor that it can hold. The vapor pressure at saturation is called the **saturation vapor pressure.** The saturation vapor pressure varies with the temperature of the air and is identical to the vapor pressure of water at that temperature. The higher the temperature, the more water vapor a volume of air can hold, and the higher the saturation vapor pressure. Conversely, the lower the temperature, the lower the saturation vapor pressure. Table 1 illustrates how the saturation vapor pressure varies with temperature. In the common range of temperatures in the lower atmosphere, the saturation vapor pressure just about doubles for each 20°F. increase in temperature. With this understanding of evaporation, condensation, and vapor pressure, we can now define several terms used to indicate the amount of moisture in the atmosphere.



The saturation absolute humidity and saturation vapor pressure both vary with the temperature. The higher the temperature, the more water vapor a volume of air can hold.

Table 1. – Saturation water vapor pressure

Temperature, °F.	Pressure, inches of mercury
-40	0.006
-30	.010
-20	.017
-10	.028
0	.045 supercooled water
10	.071
20	.110
30	.166
40	.248
50	.362
60	.522
70	.739
80	1.032
90	1.422
100	1.933
212	29.92 boiling water (sea level)

The air near the surface is usually not saturated; therefore, the actual vapor pressure is usually less than the saturation vapor pressure. The actual vapor pressure can be raised to saturation vapor pressure by evaporating more moisture into the air, or, since saturation vapor pressure varies with temperature, the air can be cooled until the saturation vapor pressure is equal to the actual vapor pressure. Evaporation alone does not ordinarily saturate the air except very close to the evaporating surface. Normal circulation usually carries evaporated moisture away from the evaporating surface.

Dew Point

Saturation is usually reached by the air being cooled until its saturation vapor pressure equals the actual vapor pressure. The temperature of the air at that point is called the dew-point temperature, or simply, the **dew point.** Further cooling causes some of the vapor to condense into liquid droplets that form clouds, fog, or dew. Cooling near the surface normally results from contact with cool ground or water. Cooling to the dew point may also occur by lifting moist air to higher altitudes; it is thus cooled adiabatically. For example, consider air with a temperature of 80°F. and a vapor pressure of 0.362 inches of mercury. Referring to table 1, we find that if the air is cooled to 500, the actual vapor pressure will equal the saturation vapor pressure. Therefore, 50° is the dew point.

If the air is cooled below its dew point, condensation occurs because the amount of water vapor in the air exceeds the maximum amount that can be contained at the lower temperature. Under ordinary circumstances the actual vapor pressure cannot exceed the saturation vapor pressure by more than a very small amount.

Absolute Humidity

The actual amount of water vapor in a given volume of air, that is, the weight per volume, such as pounds per 1,000 cubic feet, is called the **absolute humidity.** A direct relationship exists among the dew point, the vapor pressure, and the absolute humidity because, at constant atmospheric pressure, each of these depends only on the actual amount of water vapor in the air. At saturation, the dew point is the same as the temperature, the vapor pressure is the saturation vapor pressure, and the absolute humidity is the saturation absolute humidity.

Table 2 shows the relationship among these three measures of atmospheric moisture. Saturation values of vapor pressure and absolute humidity can be obtained by entering temperature

Table 2. - Dew point, vapor pressure, and absolute humidity

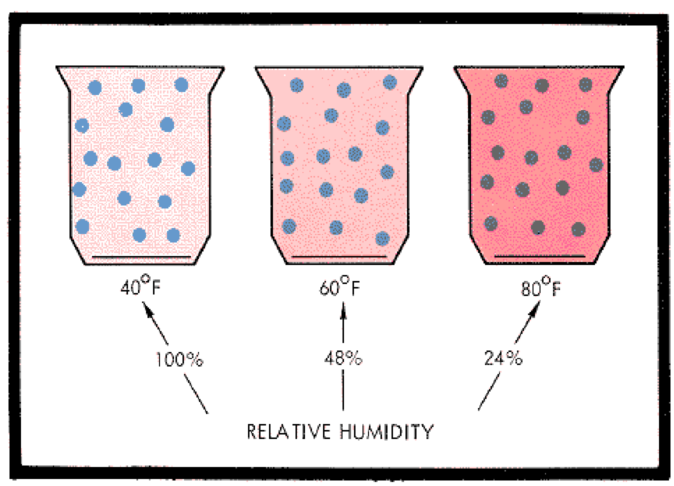
Dew point (temperature) (°F.)	Vapor pressure (saturation) (Inches of Hg.)	Absolute humidity (saturation) (Pounds per M cubic feet)
-40	0.006	0.011
-30	.010	.019
-20	.017	.031
-10	.028	.051
0	.045	.081
10	.071	.125
20	.110	.198
30	.166	.279
40	.248	.409
50	.362	.585
60	.522	.827
70	.739	1.149
80	1.032	1.575
90	1.422	2.131
100	1.933	2.844
110	2.597	3.754

instead of dew point in the first column. Because of these relationships, the temperature of the dew point is a convenient unit of measure for moisture. Air temperature and dew point accurately define atmospheric moisture at any time or place.

Relative Humidity

Saturation of surface air is a condition of favorable fire weather; that is, conducive to low fire danger. Less favorable are conditions of unsaturation, which permit evaporation from forest fuels, increasing their flammability and the fire danger. Therefore, a very useful measure of atmospheric moisture is the **relative humidity.** It is the ratio, in percent, of the amount of moisture in a volume of air to the total amount which that volume can hold at the given temperature and atmospheric pressure. Relative humidity is also the ratio of actual vapor pressure to saturation vapor pressure, times 100. It ranges from 100 percent at saturation to near zero for very dry air. Relative humidity depends on the actual moisture content of the air, the temperature, and the pressure.

The dependence of relative humidity on temperature must be kept in mind. Suppose that we have air at 800F. and 24 percent relative humidity. Using table 2, we find that the saturation vapor pressure for 800 is 1.032 inches of mercury. We can compute the actual vapor pressure by multiplying 1.032 by 0.24. The actual vapor pressure is 0.248 rounded off. The



Relative humidity decreases as temperature increases even though the amount of water vapor in the air remains the some.

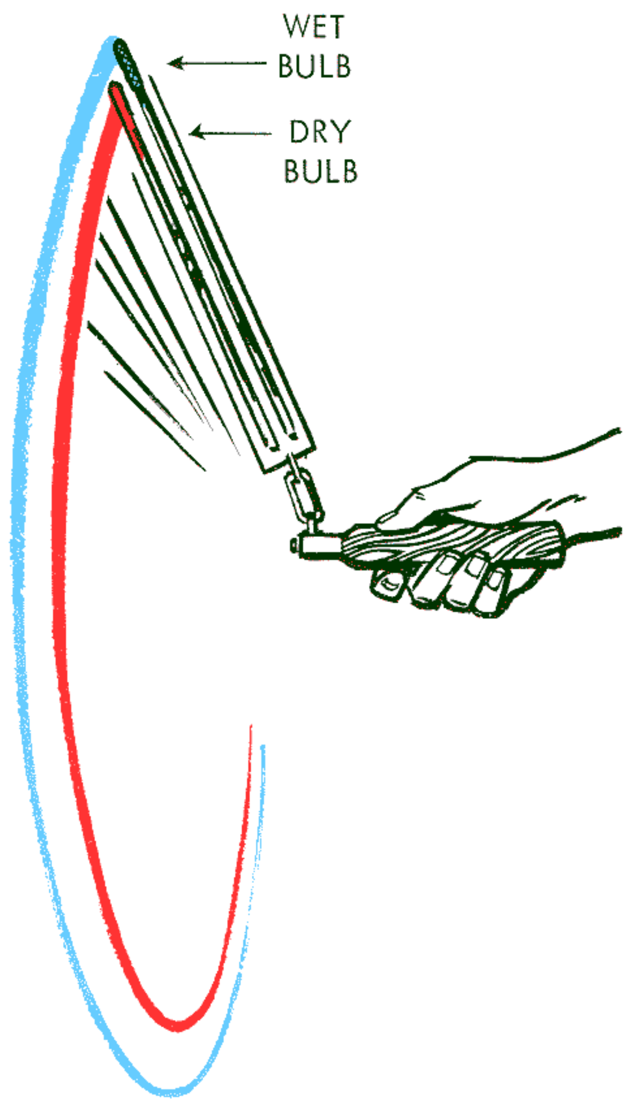
dew point for this vapor pressure is 40°. We now know that if the air was cooled from 80°F. to 40°, with no other change, the humidity would increase from 24 percent to 100 percent and the air would be saturated. At that temperature the actual vapor pressure would equal the saturation vapor pressure. The absolute humidity in table 2 could be used in a similar manner. Thus, the relative humidity may change considerably with no addition of moisture-just by cooling alone.

MEASURING HUMIDITY

The most widely used device for accurately measuring atmospheric moisture near the surface is the **psychrometer.** It consists of two identical mercurial thermometers. One thermometer is used for measuring the air temperature; the other measures the temperature of evaporating water contained in a muslin wicking surrounding the thermometer bulb. The amount that the evaporating surface will cool is determined by the difference between the vapor pressure and the saturation vapor pressure. The first reading is commonly referred to as the dry-bulb temperature and the second as the wet-bulb temperature. The wet-bulb temperature is the steady value reached during a period of brisk ventilation of the thermometer bulbs. If the air is saturated, the wet-bulb and dry-bulb temperatures are the same.

From the wet- and dry-bulb measurements, computed values of dew-point temperature, absolute humidity, and relative humidity may be read from tables or slide rules. As noted earlier, these moisture relations vary with changes in pressure. The daily pressure changes as shown by the barometer are not large enough to be important, but those due to differences in elevation are significant. They have been considered in the construction of the tables or slide rules. The ones labeled with the correct pressure must be used. Table 3 gives the ranges of land elevations for which **psychrometric tables** for different pressures may be used.

Table 4 is a sample of one of the simplest types of tables. Either relative humidity or dew point may be obtained directly from wet-bulb and dry-bulb readings. As an example, suppose the air temperature (dry-bulb) was 75°F.



Wet-bulb and dry-bulb temperatures are obtained with a psychrometer. Dew point, relative humidity, and other measures of air moisture may be obtained from these readings.

and the wet-bulb temperature was 64° at a station 1,500 feet above sea level. Entering table 4 (which is the table for 29 inches of mercury) with the dry-bulb reading on the left and the wet-bulb reading at the top, we find at the intersection that the relative humidity is 55 percent (black figure) and the dew point is 58°F. (red figure).

Other tables in common use require that the wet-bulb depression (the dry-bulb temperature minus the wet-bulb temperature) be computed first. One table is entered with this value and the dry-bulb reading to obtain the dew point; another table is entered with the same two readings to obtain the relative humidity.

Table 3. – Psychrometric tables for different Elevations

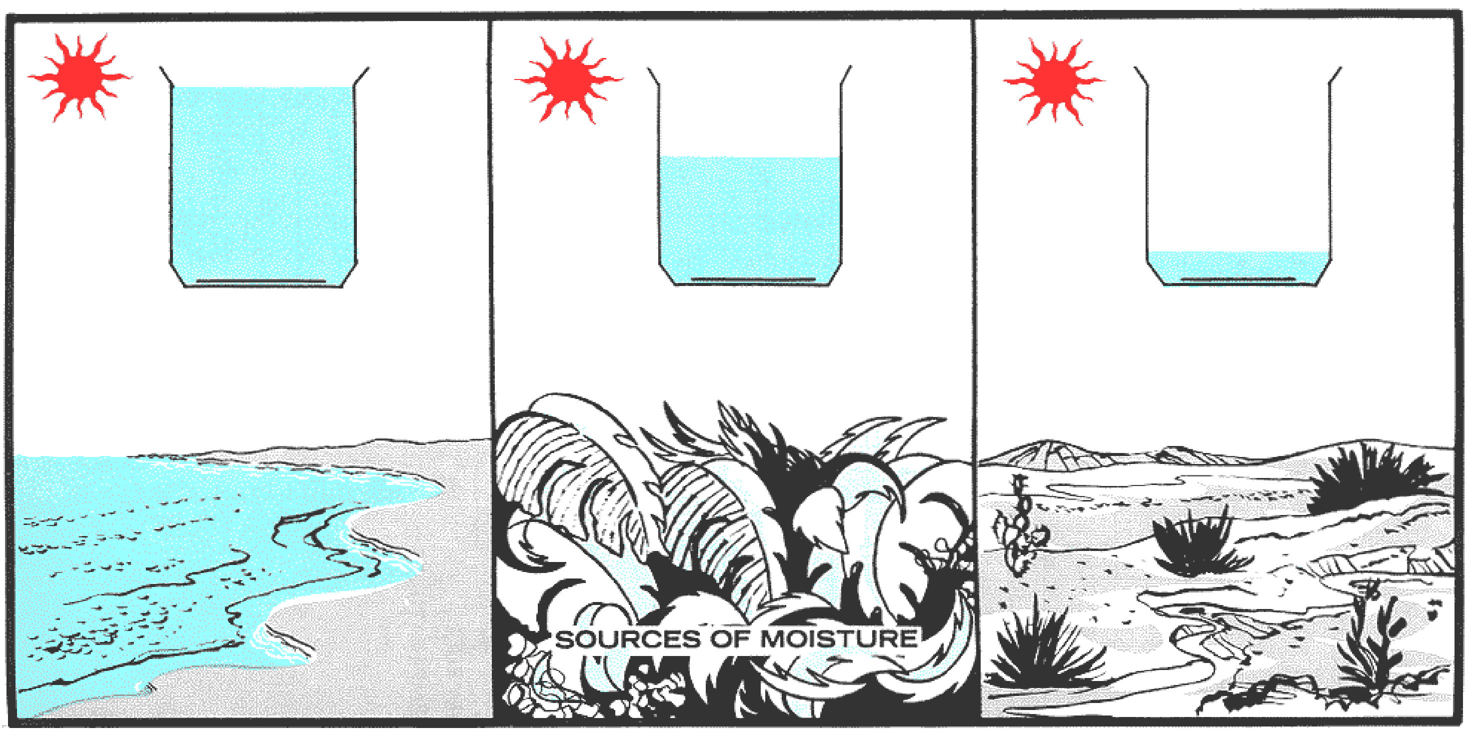
Elevation above sea level (Except Alaska) (Feet)	(Alaska)	Psychrometric table (Inches of hg.)
0-500	0-300	30
501-1900	301-1700	29
1901-3900	1701-3600	27
3901-6100	3601-5700	25
6101-8500	5701-7900	23

Other instruments used to measure relative humidity contain fibers of various materials that swell or shrink with changing relative humidity. One instrument of this type that records a continuous trace of relative humidity is called a **hygrograph.** A more common form in use at fire-weather stations is the **hygrothermograph,** which records both relative humidity and temperature. Other devices, such as those commonly used for upper-air soundings, employ moisture-sensitive elements that change in electrical or chemical characteristics with changing humidity.

Standard surface measurements of relative humidity, like those of temperature, are made in an instrument shelter 4 1/2 feet above the ground. A properly operated sting psychrometer, however, will indicate dry- and we-bulb readings that agree well with those obtained in the shelter. The only necessary precautions are to select a well-ventilated shady spot, and to whirl the instrument rapidly for a sufficient time to get the true (lowest) wet-bulb temperature. Care must be taken not to allow the wicking to dry out, and not to break the thermometer by striking any object while whirling the psychrometer.

SOURCES OF ATMOSPHERIC MOISTURE

Water vapor in the air comes almost entirely from three sources: Evaporation from any moist surface or body of water, evaporation from soil, and transpiration from plants. Some water vapor results from combustion. Because the oceans cover more than three-fourths of



Although the oceans are the principal source of atmospheric moisture, transpiration from plant, is also important. But in and areas, transpiration adds little moisture to the atmosphere.

Table 4. – Relative humidity and dew-point table for use at elevations between 501 and 1900 feet above sea level. Relative humidity in percent is shown in black: dew point in °F. is shown in red.

WET BULB TEMPERATURES

DRY BULB TEMPERATURES

(Each cell: dew point °F. / relative humidity %)

Dry bulb	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64
61	-50 / 1	-14 / 4	-1 / 8	+8 / 12	14 / 16	19 / 20	24 / 24	28 / 28	31 / 32	34 / 36	37 / 41	40 / 45	42 / 50	44 / 54	47 / 59	49 / 64	51 / 69	53 / 74	54 / 79	56 / 84	58 / 89	59 / 95	61 / 100			
62		-25 / 2	-6 / 6	+4 / 10	11 / 13	17 / 17	22 / 21	26 / 25	30 / 29	33 / 33	36 / 37	38 / 42	41 / 46	43 / 50	46 / 55	48 / 60	50 / 64	52 / 69	54 / 74	55 / 79	57 / 84	59 / 89	60 / 95	62 / 100		
63		-58	-14 / 4	-1 / 7	+8 / 11	14 / 15	20 / 19	24 / 22	28 / 26	31 / 30	34 / 34	37 / 38	40 / 43	42 / 47	45 / 51	47 / 56	49 / 60	51 / 65	53 / 70	55 / 74	57 / 79	58 / 84	60 / 89	61 / 95	63 / 100	
64			-27 / 2	-7 / 5	+4 / 9	11 / 12	17 / 16	22 / 20	26 / 24	30 / 27	33 / 31	36 / 35	39 / 39	41 / 43	44 / 48	46 / 52	48 / 56	50 / 61	52 / 65	54 / 70	56 / 75	58 / 80	59 / 85	61 / 90	62 / 95	64 / 100
65				-15 / 3	-1 / 7	+8 / 10	14 / 14	20 / 17	24 / 21	28 / 25	32 / 29	35 / 32	38 / 36	40 / 40	43 / 44	45 / 48	47 / 53	49 / 57	51 / 61	53 / 66	55 / 70	57 / 75	59 / 80	60 / 85	62 / 90	63 / 95
66				-28 / 2	-7 / 5	+4 / 8	11 / 12	17 / 15	22 / 19	26 / 22	30 / 26	33 / 30	36 / 33	39 / 37	42 / 41	44 / 45	46 / 49	49 / 53	51 / 58	53 / 62	54 / 66	56 / 71	58 / 76	60 / 80	61 / 85	63 / 90
67					-15 / 3	-1 / 6	+8 / 10	15 / 13	20 / 16	24 / 20	28 / 23	32 / 27	35 / 31	38 / 34	41 / 38	43 / 42	46 / 46	48 / 50	50 / 54	52 / 58	54 / 63	56 / 67	57 / 71	59 / 76	61 / 80	62 / 85
68					-29 / 2	-7 / 5	+4 / 8	12 / 11	18 / 14	22 / 18	27 / 21	30 / 24	34 / 28	37 / 32	40 / 35	42 / 39	45 / 43	47 / 47	49 / 51	51 / 55	53 / 59	55 / 63	57 / 67	59 / 72	60 / 76	62 / 81
69						-15 / 3	-1 / 6	+8 / 9	15 / 12	20 / 16	25 / 19	29 / 22	32 / 26	35 / 29	38 / 33	41 / 36	44 / 40	46 / 44	48 / 47	50 / 51	52 / 55	54 / 59	56 / 64	58 / 68	60 / 72	61 / 77
70						-29 / 1	-7 / 4	+4 / 7	12 / 10	18 / 14	23 / 17	27 / 20	31 / 23	34 / 27	37 / 30	40 / 34	43 / 37	45 / 41	47 / 44	50 / 48	52 / 52	54 / 56	55 / 60	57 / 64	59 / 68	61 / 72
71							-15 / 3	0 / 6	+9 / 9	15 / 12	21 / 15	25 / 18	29 / 21	33 / 24	36 / 28	39 / 31	42 / 34	44 / 38	46 / 42	49 / 45	51 / 49	53 / 53	55 / 57	57 / 60	58 / 65	60 / 69
72							-28 / 1	-6 / 4	+5 / 7	12 / 10	18 / 13	23 / 16	28 / 19	31 / 22	35 / 25	38 / 29	40 / 32	43 / 35	46 / 39	48 / 42	50 / 46	52 / 50	54 / 53	56 / 57	58 / 61	60 / 65
73								-14 / 3	0 / 5	+9 / 8	16 / 11	21 / 14	26 / 17	30 / 20	33 / 23	36 / 26	39 / 30	42 / 33	45 / 36	47 / 40	49 / 43	51 / 47	53 / 50	55 / 55	57 / 58	59 / 62
74								-27 / 1	-6 / 4	+5 / 7	13 / 10	19 / 12	24 / 15	28 / 18	32 / 21	35 / 24	38 / 27	41 / 30	44 / 34	46 / 37	48 / 40	51 / 44	53 / 47	55 / 51	57 / 54	58 / 58
75									-13 / 3	+1 / 5	10 / 8	16 / 11	22 / 13	26 / 16	30 / 19	34 / 22	37 / 25	40 / 28	43 / 31	45 / 35	48 / 38	50 / 41	52 / 44	54 / 48	56 / 51	58 / 55
76									-25 / 1	-5 / 4	+6 / 7	14 / 9	20 / 12	24 / 15	29 / 17	32 / 20	36 / 23	39 / 26	42 / 29	44 / 32	47 / 35	49 / 39	51 / 42	53 / 45	55 / 49	57 / 52
77									-57	-12 / 3	+2 / 5	10 / 8	17 / 10	22 / 13	27 / 16	31 / 18	34 / 21	38 / 24	41 / 27	43 / 30	46 / 33	48 / 36	50 / 39	53 / 43	55 / 46	57 / 49
78										-23 / 1	-4 / 4	+7 / 6	14 / 9	20 / 11	25 / 14	29 / 17	33 / 19	36 / 22	39 / 25	42 / 28	45 / 31	47 / 34	50 / 37	52 / 40	54 / 43	56 / 46
79										-48	-11 / 3	+3 / 5	11 / 8	18 / 10	23 / 13	28 / 15	32 / 18	35 / 20	38 / 23	41 / 26	44 / 29	46 / 32	49 / 35	51 / 38	53 / 41	55 / 44
80											-21 / 2	-2 / 4	+8 / 6	15 / 9	21 / 11	26 / 14	30 / 16	34 / 19	37 / 21	40 / 24	43 / 27	45 / 30	48 / 32	50 / 35	52 / 38	54 / 41
81											-40 / 1	-9 / 3	+4 / 5	12 / 7	19 / 10	24 / 12	28 / 15	32 / 17	36 / 20	39 / 22	42 / 25	45 / 28	47 / 30	49 / 33	52 / 36	54 / 39
82												-18 / 2	-1 / 4	+9 / 6	16 / 8	22 / 10	27 / 13	31 / 16	34 / 18	38 / 21	41 / 23	44 / 26	46 / 28	49 / 31	51 / 34	53 / 37
83												-34 / 1	-7 / 3	+5 / 5	13 / 7	20 / 9	25 / 12	29 / 14	33 / 17	36 / 19	40 / 21	43 / 24	45 / 27	48 / 29	50 / 32	52 / 35
84													-15 / 2	+1 / 4	10 / 6	17 / 8	23 / 10	27 / 13	32 / 15	35 / 17	38 / 20	41 / 22	44 / 25	47 / 27	49 / 30	52 / 33
85													-29 / 1	-5 / 3	+7 / 5	14 / 7	21 / 9	26 / 11	30 / 14	34 / 16	37 / 18	40 / 21	43 / 23	46 / 26	48 / 28	51 / 31
86														-12 / 2	+2 / 4	11 / 6	18 / 8	24 / 10	28 / 12	32 / 15	36 / 17	39 / 19	42 / 22	45 / 24	48 / 26	50 / 29
87														-24 / 1	-3 / 3	+8 / 5	16 / 7	22 / 9	27 / 11	31 / 13	35 / 16	38 / 18	41 / 20	44 / 22	47 / 25	49 / 27
88														-50	-10 / 2	+4 / 4	13 / 6	19 / 8	25 / 10	29 / 12	33 / 14	37 / 16	40 / 19	43 / 21	46 / 23	48 / 26
89															-19 / 1	-1 / 3	+10 / 5	17 / 7	23 / 9	28 / 11	32 / 13	36 / 15	39 / 17	42 / 19	45 / 22	48 / 24
90															-37	-7 / 2	+5 / 4	14 / 6	21 / 8	26 / 10	30 / 12	34 / 14	38 / 16	41 / 18	44 / 20	47 / 22

Dry bulb	65	66	67	68	69	70	71	72	73	74	75	76	77	78	79	80	81	82	83	84	85	86	87	88	89	90
65	65 / 100																									
66	64 / 95	66 / 100																								
67	64 / 90	66 / 95	67 / 100																							
68	63 / 85	65 / 90	67 / 95	68 / 100																						
69	63 / 81	65 / 86	66 / 90	68 / 95	69 / 100																					
70	62 / 77	64 / 81	66 / 86	67 / 90	69 / 95	70 / 100																				
71	62 / 73	63 / 77	65 / 82	67 / 86	68 / 91	70 / 95	71 / 100																			
72	61 / 69	63 / 73	65 / 77	66 / 82	68 / 86	69 / 91	71 / 95	72 / 100																		
73	61 / 65	62 / 69	64 / 74	66 / 78	67 / 82	69 / 86	70 / 91	72 / 95	73 / 100																	
74	60 / 62	62 / 66	64 / 70	65 / 74	67 / 78	68 / 82	70 / 87	71 / 91	73 / 95	74 / 100																
75	60 / 59	61 / 62	63 / 66	65 / 70	66 / 74	68 / 78	69 / 82	71 / 87	72 / 91	74 / 95	75 / 100															
76	59 / 56	61 / 59	62 / 63	64 / 67	66 / 71	67 / 75	69 / 79	70 / 83	72 / 87	73 / 91	75 / 96	76 / 100														
77	58 / 53	60 / 56	62 / 60	64 / 63	65 / 67	67 / 71	68 / 75	70 / 79	71 / 83	73 / 87	74 / 91	76 / 96	77 / 100													
78	58 / 50	60 / 53	61 / 57	63 / 60	65 / 64	66 / 67	68 / 71	69 / 75	71 / 79	72 / 83	74 / 87	75 / 91	77 / 96	78 / 100												
79	57 / 47	59 / 50	61 / 54	63 / 57	64 / 61	66 / 64	67 / 68	69 / 72	71 / 75	72 / 79	73 / 83	75 / 87	76 / 91	78 / 96	79 / 100											
80	56 / 45	58 / 48	60 / 51	62 / 54	64 / 58	65 / 61	67 / 65	69 / 68	70 / 72	72 / 76	73 / 79	75 / 83	76 / 87	77 / 92	79 / 96	80 / 100										
81	56 / 42	58 / 45	60 / 48	61 / 51	63 / 55	65 / 58	67 / 62	68 / 65	70 / 69	71 / 72	73 / 76	74 / 80	76 / 84	77 / 88	78 / 92	80 / 96	81 / 100									
82	55 / 40	57 / 43	59 / 46	61 / 49	63 / 52	64 / 55	66 / 59	68 / 62	69 / 65	71 / 69	72 / 73	74 / 76	75 / 80	77 / 84	78 / 88	79 / 92	81 / 96	82 / 100								
83	54 / 38	56 / 40	58 / 43	60 / 46	62 / 49	64 / 53	66 / 56	67 / 59	69 / 62	70 / 66	72 / 69	73 / 73	75 / 76	76 / 80	78 / 84	79 / 88	80 / 92	82 / 96	83 / 100							
84	54 / 35	56 / 38	58 / 41	60 / 44	62 / 47	63 / 50	65 / 53	67 / 56	68 / 59	70 / 63	71 / 66	73 / 70	74 / 73	76 / 77	77 / 80	79 / 84	80 / 88	81 / 92	83 / 96	84 / 100						
85	53 / 33	55 / 36	57 / 39	59 / 42	61 / 45	63 / 47	65 / 50	66 / 54	68 / 57	69 / 60	71 / 63	73 / 66	74 / 70	76 / 73	77 / 77	78 / 81	80 / 84	81 / 88	82 / 92	84 / 96	85 / 100					
86	52 / 32	54 / 34	57 / 37	59 / 40	60 / 42	62 / 45	64 / 48	66 / 51	67 / 54	69 / 57	71 / 60	72 / 64	74 / 67	75 / 70	77 / 74	78 / 77	79 / 81	81 / 84	82 / 88	83 / 92	85 / 96	86 / 100				
87	52 / 30	54 / 32	56 / 35	58 / 37	60 / 40	62 / 43	64 / 46	65 / 49	67 / 51	69 / 54	70 / 58	72 / 61	73 / 64	75 / 67	76 / 70	78 / 74	79 / 77	80 / 81	82 / 85	83 / 88	84 / 92	86 / 96	87 / 100			
88	51 / 28	53 / 30	55 / 33	57 / 35	59 / 38	61 / 41	63 / 43	65 / 46	66 / 49	68 / 52	70 / 55	71 / 58	73 / 61	74 / 64	76 / 67	77 / 71	79 / 74	80 / 78	81 / 81	83 / 85	84 / 88	85 / 92	87 / 96	88 / 100		
89	50 / 26	52 / 29	55 / 31	57 / 34	59 / 36	61 / 39	62 / 41	64 / 44	66 / 47	68 / 50	69 / 52	71 / 55	72 / 58	74 / 61	76 / 65	77 / 68	78 / 71	80 / 74	81 / 78	83 / 81	84 / 85	85 / 88	86 / 92	88 / 96	89 / 100	
90	49 / 25	52 / 27	54 / 29	56 / 32	58 / 34	60 / 37	62 / 39	64 / 42	66 / 45	67 / 47	69 / 50	71 / 53	72 / 56	74 / 59	75 / 62	77 / 65	78 / 68	79 / 71	81 / 75	82 / 78	84 / 82	85 / 85	86 / 89	87 / 92	89 / 96	90 / 100

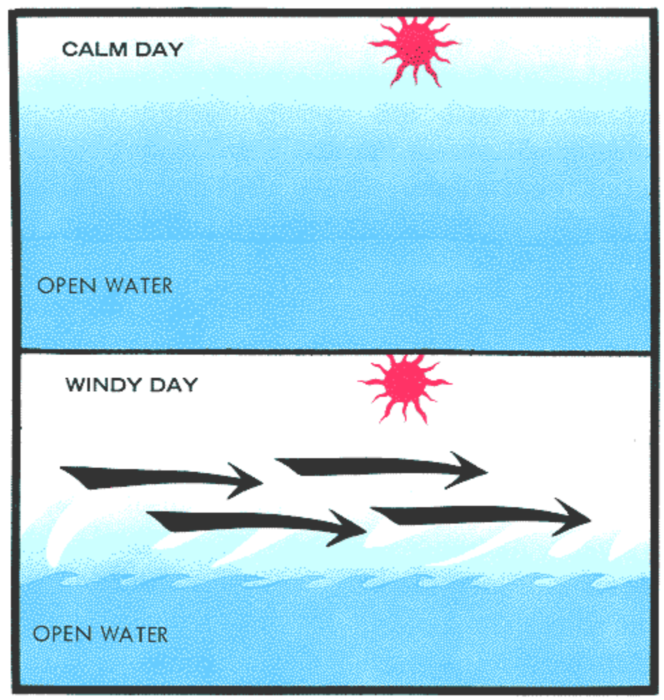
the earth's surface, they are the most important moisture source, but land sources can also be important locally.

Plants have large surfaces for transpiration; occasionally they have as much as 40 square yards for each square yard of ground area. Transpiration from an area of dense vegetation can contribute up to eight times as much moisture to the atmosphere as can an equal area of bare ground. The amount of moisture transpired depends greatly on the growth activity. This growth activity, in turn, usually varies with the season and with the ground water supply. In areas of deficient rainfall and sparse vegetation, such as many areas in the arid West, both transpiration and evaporation may be almost negligible toward the end of the dry season. This may also be common at timberline and at latitudes in the Far North.

In evaporation from water bodies, soil, and dead plant material, the rate at which moisture is given up to the air varies with the difference between the vapor pressure at the evaporating surface and the atmospheric vapor pressure. Evaporation will continue as long as the vapor pressure at the evaporating surface is greater than the atmospheric vapor pressure. The rate of evaporation increases with increases in the pressure difference. The vapor pressure at the evaporating surface varies with the temperature of that surface. Therefore, evaporation from the surfaces of warm water bodies, warm soil, and dead plant material will be greater than from cold surfaces, assuming that the atmospheric vapor pressure is the same.

Transpiration from living plants does not vary as evaporation from dead plant material. Living plants will usually transpire at their highest rates during warm weather, but an internal regulating process tends to limit the water-loss rate on excessively hot and dry days to the plant's particular current needs. We will discuss evaporation from dead **plant material** and transpiration from living plants more fully in the chapter on fuel moisture (chapter 11).

In still air during evaporation, water vapor concentrates near the evaporating surface. If this concentration approaches saturation, further evaporation will virtually halt, even though the surrounding air is relatively dry. **Wind** encourages evaporation by blowing away these stagnated layers and replacing them with drier air. After a surface has dried to the point where free water is no longer exposed to the air, the effect of wind on evaporation decreases. In fact, for surfaces like comparatively dry soil or wood, wind may actually help reverse the process by cooling the surfaces and thus lowering the vapor pressure of moisture which these surfaces contain.



Wind encourages evaporation by blowing away stagnated layers of moist air and by mixing moist air with drier air aloft.

VARIATIONS IN ABSOLUTE HUMIDITY

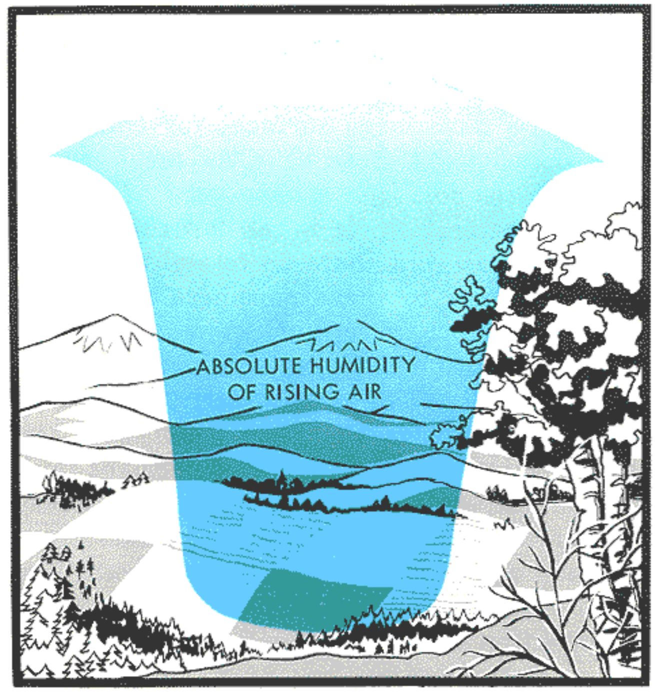
The actual amount of moisture in the air will vary from one air rental to another, and even within an air mass there will be continuing variations in time and space.

The moisture contents of air masses are basically related to their regions of origin. Air masses originating in continental areas are relatively dry. Those coming from the Atlantic or the Gulf of Mexico are moist, and those from the Pacific are moist or moderately moist. As these maritime air masses invade the continent, land stations will observe abrupt rises in absolute humidity. As any air mass traverses areas different from its source region, gradual changes take place as evaporation, transpiration, condensation, and precipitation add or subtract moisture.

Through a deep layer within an air mass, the absolute humidity, like the temperature, usually **decreases with height.** There are several reasons for this distribution. **First**, moisture is added to the atmosphere from the surface and is carried upward by convection and upslope and up valley winds. **Second,** when air is lifted, the water vapor, as well as the air, expands proportionately so that the moisture in any given volume becomes less and less. Thus, the absolute humidity decreases as the air is lifted. **Third,** since temperature usually decreases upward, the capacity for air to hold moisture decreases upward. **Finally,** the precipitation process removes condensed moisture from higher levels in the atmosphere and deposits it at the surface.

The normal pattern of decrease of moisture with altitude may be altered occasionally when horizontal flow at intermediate levels aloft brings in moist air. Such flow is responsible for much of the summer thunderstorm activity over large parts of the West. Extremely low absolute humidity is found in subsiding air aloft. This dry air originates near the top of the troposphere and slowly sinks to lower levels. If it reaches the ground, or is mixed downward, it may produce acutely low humidity near the surface and an abrupt increase in fire danger. We will consider **subsidence** in more detail in the next chapter.

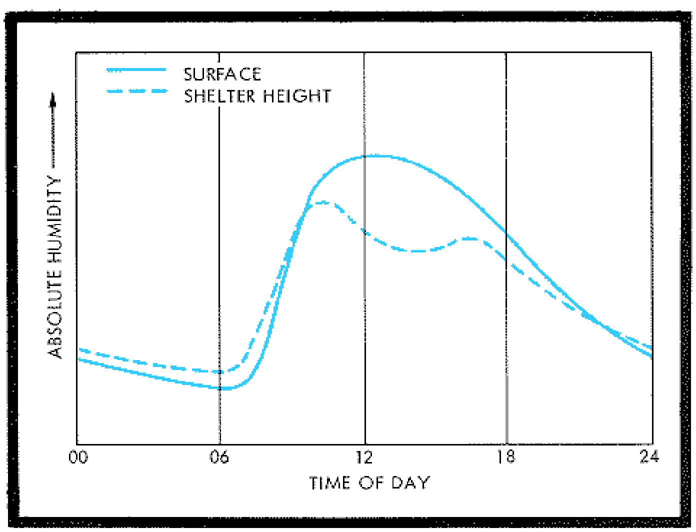
If we consider only a very shallow layer of air near the surface, we find that the vertical variation of absolute humidity with height will change during each 24-hour period as conditions favoring evaporation alternate with conditions favoring condensation. During clear days, moisture usually is added to the air by evaporation



As moist air rises, it expands, and the moisture in a given volume, the absolute humidity, becomes less and less.

from warm surfaces; therefore, the absolute humidity decreases upward.

At night, moisture is usually taken from the air near the surface by condensation on cold surfaces and absorption by cold soil and other substances; thus, the absolute humidity may increase upward through a very shallow layer.

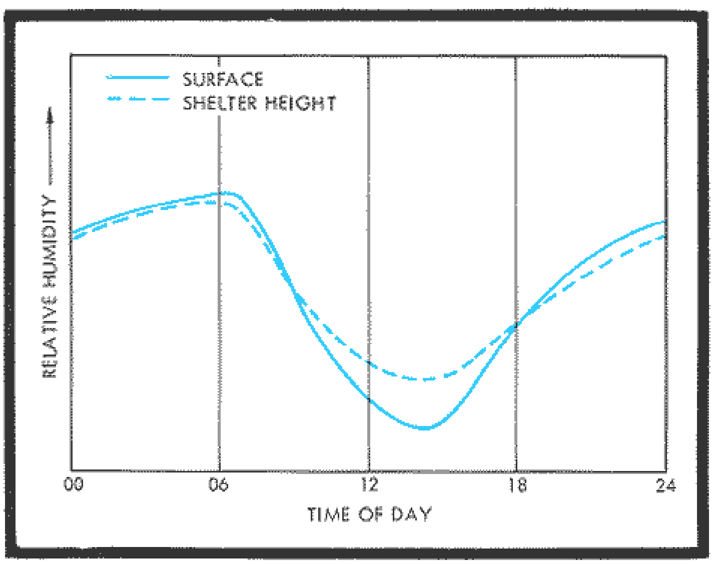


Schematic representation of surface absolute humidity compared to that at shelter height. Air near the surface is likely to contain less moisture than air at shelter height during the night, and more moisture during the day.

DIURNAL AND SEASONAL CHANGES IN RELATIVE HUMIDITY

Relative humidity is much more variable than absolute humidity. It often changes rapidly and in significant amounts from one hour to the next and from place to place. Relative humidity is much more variable because it depends not only on absolute humidity but also on air temperature. It varies directly with moisture content and inversely with temperature. Because of these relationships, it is often not possible to make general statements about relative humidity variations, particularly vertical variations within short distances above the ground.

During the day near the surface, particularly with clear skies, both the temperature and absolute humidity usually decrease with height. These two variables have opposite effects on the relative humidity. Which effect is dominant depends upon the dryness of the surface. The relative humidity usually increases with height over normal surfaces because the effect of the decrease in temperature is greater than that of the decrease in absolute humidity. Over a moist surface, however, the effect of the decrease in absolute humidity may overbalance that of temperature decrease, and the relative humidity in the surface layer will decrease with height.



Schematic representation of surface relative humidity compared to that at shelter height. Due to the effect of temperature, relative humidity near the ground is usually lower than at shelter height during the day, and higher at night.

At night, the change of temperature with height usually predominates, and the relative humidity will decrease with height through the lowest layers.

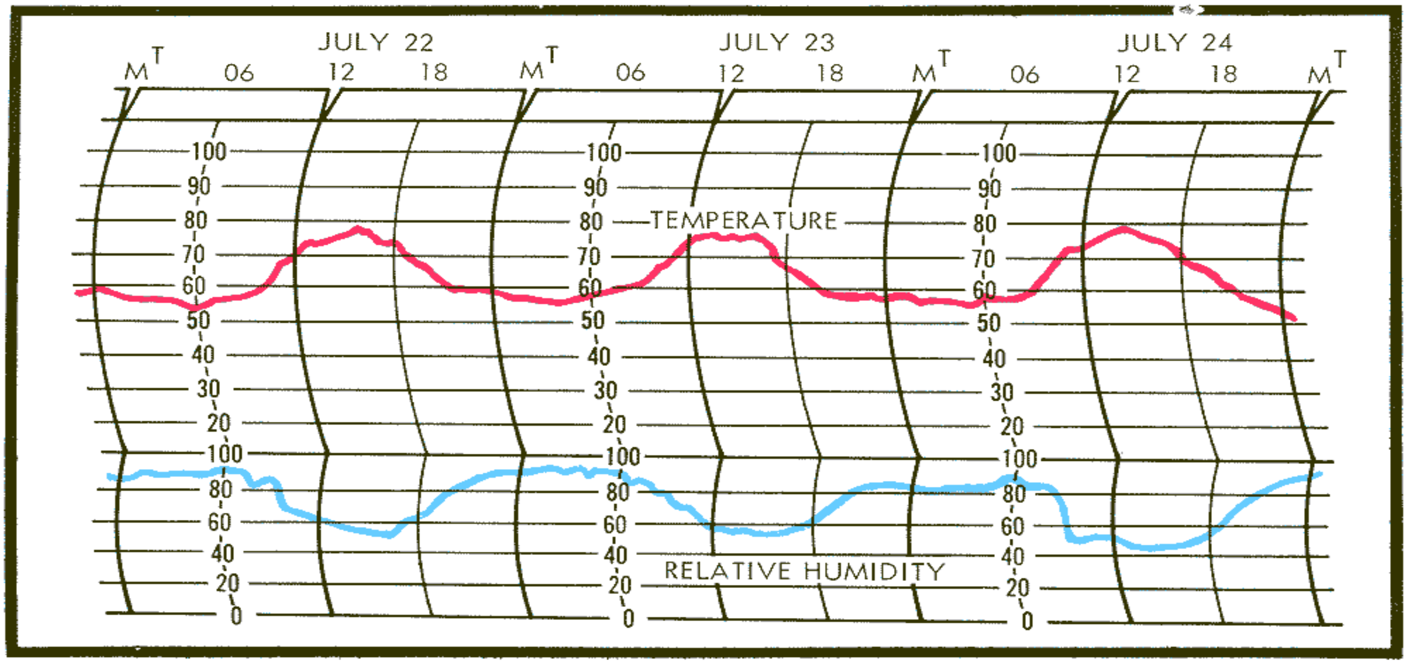
Above the lowest layers, the relative humidity generally increases with height in the day through much of the lower troposphere. Convection alone would account for this increase. As air is lifted, the temperature decreases 5.5°F. per 1,000 feet, and the dew point decreases at about 1°F. per 1,000 feet. Therefore, the dew point and the temperature become 4.5°F. closer per 1,000 feet, and the relative humidity increases until saturation is reached.

A subsiding layer of air in the troposphere warms by the adiabatic process and forms a **subsidence inversion**. The relative humidity will decrease upward through the temperature inversion at the base of the subsiding layer. The marine inversion along the west coast, for example, is a subsidence inversion. The marine air below has low temperatures and high humidities, and the adiabatically heated subsiding air mass above has higher temperatures and lower humidities. This pronounced change in temperature and humidity is evident along the slopes of coastal mountains when the marine inversion is present.

Relative humidity is most important as a fire-weather factor in the layer near the ground, where it influences both fuels and fire behavior. Near the ground, air moisture content, season, time of day, slope, aspect, elevation, clouds, and vegetation all cause important variations in relative humidity.

Since hourly and daily changes of relative humidity are normally measured in a standard instrument shelter, we will consider variations at that level and infer from our knowledge of surface temperatures what the conditions are near the surface around forest fuels.

A typical fair-weather pattern of relative humidity, as shown on a hygrothermograph exposed in a shelter at a valley station or one in flat terrain, is nearly a mirror image of the temperature pattern. Maximum humidity generally occurs about daybreak, at the time of



Typical temperature and relative humidity traces for a low-level station are nearly mirror images of each other.

minimum temperature. After sunrise, humidity drops rapidly and reaches a minimum at about the time of maximum temperature. It rises more gradually from late afternoon through the night. The daily range of humidity is usually greatest when the daily range of temperature is greatest. Variations in the humidity traces within an air mass from one day to the next are usually small, reflecting mostly differences in temperatures. But over several days, there may be noticeable cumulative differences in humidity as the air mass gradually picks up or loses moisture.

Seasonal changes in relative humidity patterns are also apparent. In western fire-weather seasons that begin following a moist spring and continue through the summer and early fall, a seasonal change is particularly noticeable. Daily temperature ranges are greatest early in the fire season when the sun is nearly overhead and night skies are clear. Strong nighttime cooling, in combination with ample moisture in the soil and vegetation to contribute moisture to the atmosphere, often boosts night humidities to or near 100 percent. Intensive daytime surface heating and convective transport of moisture upward combine to drop the relative humidity to low levels in the afternoon.

As the season progresses, soil and vegetation dry out and solar heating diminishes as the sun tracks farther south. Daytime humidities become even lower late in the season, but, with a greater reduction in night humidities, the daily range is reduced, and the fire weather is further intensified. Occasional summer rains may interrupt this progression but do not greatly change the overall seasonal pattern.

In areas that have separate spring and fall fire seasons, the daily temperature extremes are generally not so striking. Also, the cumulative drying of soil and vegetation is not so consistent, except during unusual drought. Because periodic rains generally occur during the seasons, the humidity changes tend to be somewhat variable. In some areas, seasonal increases in relative humidity decrease fire danger during the summer. In the Great Lakes region, particularly, where the many small lakes become quite warm during the summer and transpiration from vegetation is at its peak, daytime relative humidities do not reach as low values in the same air mass types as they do in spring and fall.

The relative humidity that affects fuels on the forest floor is of ten quite different from that in the instrument shelter, particularly in unshaded areas where soil and surface fuels exposed to the sun are heated intensely, and warm the air surrounding them. This very warm air may have a dew point nearly the same or slightly higher than the air in the instrument shelter, but because it is much warmer, it has a much lower relative humidity.

It is impractical to measure humidity close to the ground with field instruments, but with the aid of tables, the humidity can be estimated from psychrometric readings at the standard height and a dry-bulb temperature reading at the surface. We must assume that the clew point is the same at both levels. Although we know that this may not be exact, it will give a reasonable estimation.

Consider the following example, using table 4, for a pressure of 29 inches:

Height of Measurement	Dry-bulb	Wet-bulb	Dew point	Relative humidity, percent
4 ½ feet	[1]80	[1]65	[2]56	[2]45
1 inch	[1]140		[3]56	[3]8

[1]Observed. [2]Calculated. [3]Estimated

The 8-percent relative was obtained from a complete set of tables, using a dry-bulb temperature of 140°F. and a dew point of 56°F.

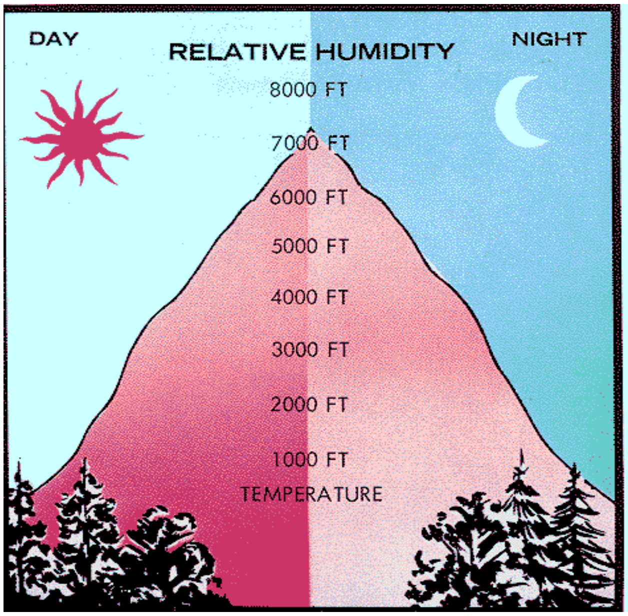
With similar exposure at night, humidities are likely to be higher near the ground than in the shelter because of radiative cooling of the surface. Often, dew will form on the surface - indicating 100 percent relative humidity-when the humidity at shelter height may be considerably below the saturation level.

These conditions are typical for relatively still air, clear skies, and open exposure. When wind speeds reach about 8 miles per hour, the increased mixing diminishes the difference between surface and shelter-height humidities. Also, under heavy cloud cover or shade, the humidity differences between the two levels tend to disappear because the principal radiating surface is above both levels.

EFFECTS OF TERRAIN, WIND, CLOUDS, VEGETATION, AND AIR MASS CHANGES

Humidity may vary considerably from one spot to another, depending greatly on the topography. In relatively flat to rolling terrain, the humidity measured at a well-exposed station may be quite representative of a fairly large area. There will be local exceptions along streams, irrigated fields, in shaded woods, or in barren areas. In the daytime particularly, circulation and mixing are usually sufficient to smooth out local effects over relatively short distances.

In mountainous topography, the effects of elevation and aspect become important, and humidities vary more than over gentle terrain. Low elevations warm up and dry out earlier in the spring than do high elevations. South slopes also are more advanced seasonally than north slopes. As the season progresses, cumulative drying tends to even out these differences since stored moisture in the surface is depleted, but the differences do not disappear.

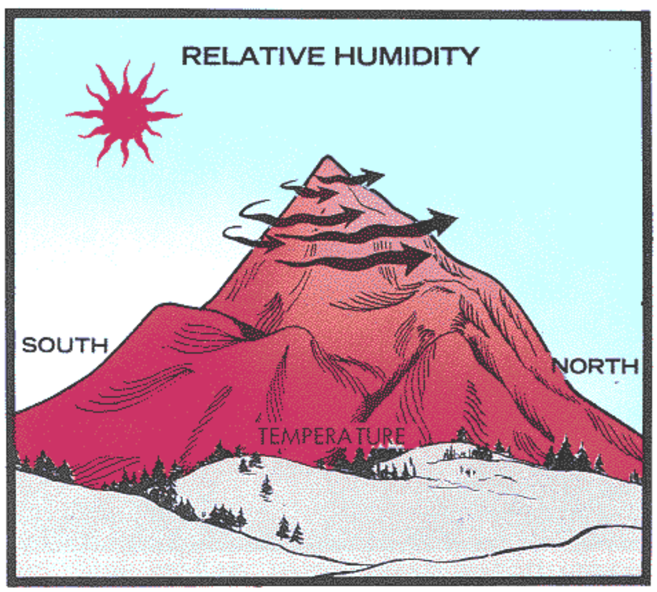


During daytime, relative humidity usually increases upward along slopes, largely because of the temperature decreases. At night, if an inversion is present, relative humidity decreases up the slope to the top of the inversion, then changes little or increases slightly with elevation.

We mentioned earlier that daytime temperatures normally decrease with altitude in the free air. The decrease with height of both temperature and dew point produces higher relative humidities at higher elevations on slopes. The pattern is complicated, however, because of heating of the air next to the slopes, the transport of moisture with upslope winds, and the frequent stratification of moisture into layers, so generalizations are difficult to make.

When nighttime cooling begins, the temperature change with height is usually reversed. Cold air flowing down the slopes accumulates at the bottom. As the night progresses, additional cooling occurs, and by morning, if the air becomes saturated, fog or dew forms. Relative humidity may decrease from 100 percent at the foot of the slope to a minimum value at the top of the temperature inversion-the thermal belt, which was discussed in chapter 2 - and then may increase slightly farther up the slope above the inversion.

Just as south slopes dry out faster because of their higher day temperatures, they also have somewhat lower day relative humidities than north slopes throughout the summer. At upper elevations, though, the difference between north and south slopes becomes negligible because of the good air mixing at these more exposed sites. At night, humidity differences on north and south slopes become slight.



During the day, south slopes have lower relative humidities than north slopes; but at upper elevations, because of good air mixing, the difference in negligible.

In most mountainous country, the daily range of relative humidity is greatest in valley bottoms and least at higher elevations. Thus, while fires on lower slopes may burn better during the day, they often quiet down considerably at night when humidity increases. But at higher elevations, particularly in and above the thermal belt, fires may continue to burn aggressively through the night as humidities remain low, temperatures stay higher, and wind speed is greater.

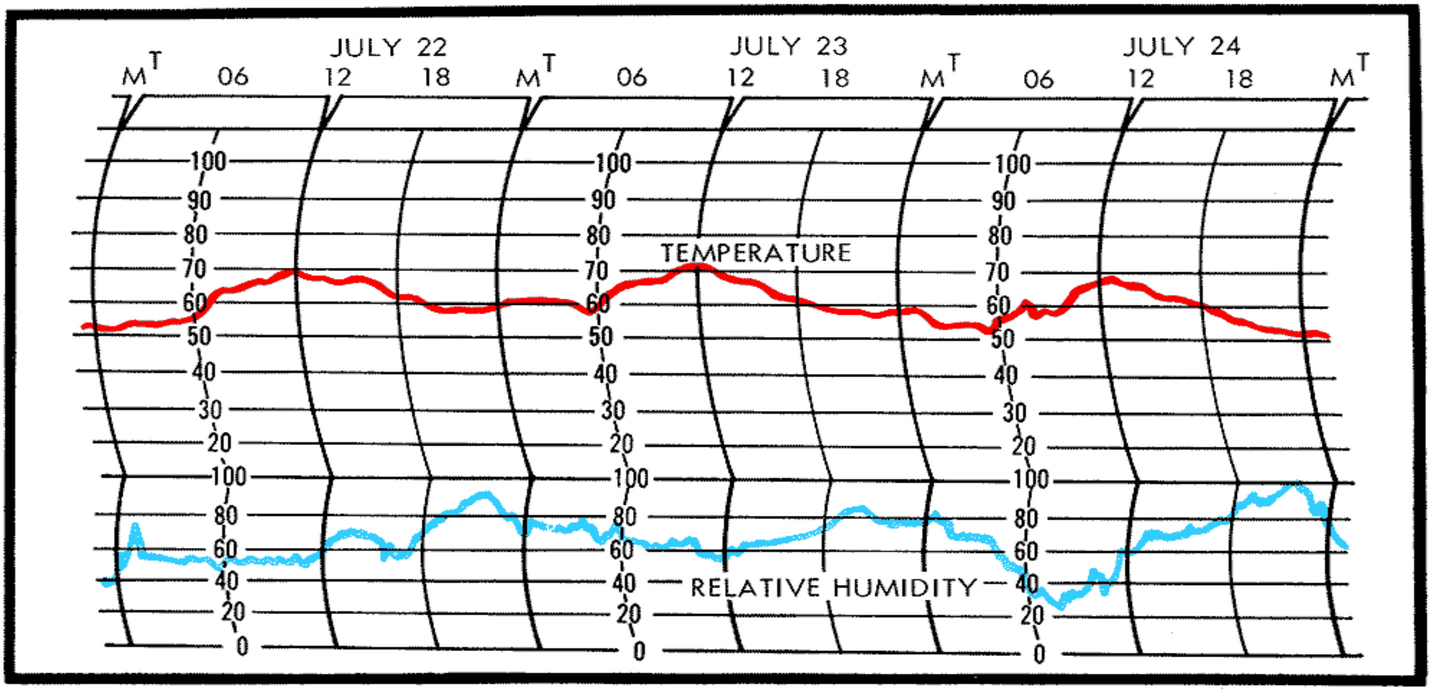
Again, we should be cautious of generalizations. For example, in the summer in the Pacific coast ranges, higher humidities are usually found on ridge tops during the day than during the night. This anomaly results from slope winds carrying moisture upward from the moist marine air layer during the day. Moist air that is not carried away aloft settles back down at night.

Wind mixes evaporating water vapor with surrounding air and evens out temperature extremes by moving air away from hot and cold surfaces. Thus, diurnal ranges of relative humidity are less during windy periods than during calm periods. Winds also reduce place-to-place differences by mixing air of different moisture contents and different temperatures. Patches of fog on a calm night indicate poor ventilation.

Clouds strongly affect heating and cooling and therefore influence the relative humidity. The humidity will be higher on cloudy days and lower on cloudy nights. Thus, clouds reduce the daily range considerably. Precipitation in any form raises relative humidities by cooling the air and by supplying moisture for evaporation into the air.

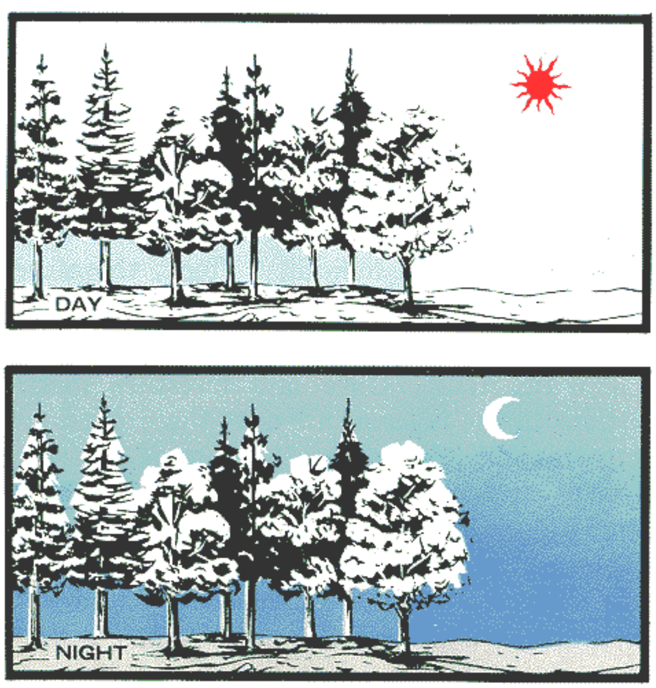
Vegetation moderates surface temperatures and contributes to air moisture through transpiration and evaporation-both factors that affect local relative humidity. A continuous forest canopy has the added effect of decreasing surface wind speeds and the mixing that takes place with air movement.

The differences in humidity between forest stands and open areas generally vary with the density of the crown canopy. Under a closed canopy, humidity is normally higher than outside during the day, and lower at night. The



Temperature and relative humidity traces at mountain stations are often less closely related ban at valley stations. Changes in absolute humidity are more important at mountain stations.

higher daytime humidities are even more pronounced when there is a green understory. Deciduous forests have only slight effects on humidity during their leafless period.



Relative humidity is normally higher under a closed canopy than in the open during the day, and lower at night.

Two factors lessen the humidity difference between forest stands and forest openings. Overcast skies limit both heating and cooling, and drought conditions decrease the amount of moisture available for evaporation and transpiration.

Openings of up to about 20 yards in diameter do not have daytime relative humidities much different from under the canopy-except at the heated ground surface. As mentioned in the previous chapter, these openings serve as chimneys for convective airflow, and surface air is drawn into them from the surrounding forest. At night in small openings, the stagnation coupled with strong radiation can cause locally high humidities.

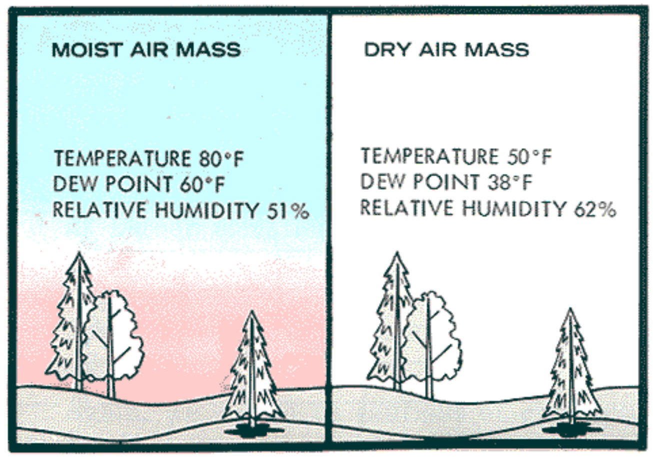
The daytime humidities in larger clearings are much like those in open country. If the airflow is restricted, however, temperatures may rise slightly above those at exposed stations, and humidities will be correspondingly lower. In the afternoon, these may range from 5 to 20 percent lower in the clearing than within a well-shaded forest. Night humidities are generally similar to those at exposed sites, usually somewhat higher than in the woods.

Open forest stands have humidity characteristics somewhere between those of exposed sites and closed stands, depending on crown density. During dry weather, especially after prolonged dry spells, the differences in relative humidity between forested and open lands become progressively less.

This discussion of relative humidity variations has so far considered changes only within an air mass. As we will see later in the chapter on air masses and fronts, the amount of moisture in the air is one of the air-mass characteristics. Air masses originating over water bodies will have higher moisture contents than those originating over continents.

When a front passes, and a different air mass arrives, a change in absolute humidity can be expected. The change in relative humidity, however, will depend greatly on the air-mass temperature. A warm, dry air mass replacing a cool, moist one, or vice versa, may cause a large change in relative humidity. A cool, dry air mass replacing a warm, moist one, however, may actually have a higher relative humidity if its temperature is appreciably lower.

Along the west coast, when a lower marine layer is topped by a warm, dry, subsiding air mass, the inversion layer is actually the boundary between two very different air masses. Inland, where the inversion intersects the



A cool, dry air mass may actually have a higher relative humidity than a warm, moist air mass.

coast ranges, very abnormal relative humidity patterns are found. In these inland areas, the inversion is usually higher in the day and lower at night; however, along the coastal lowlands, the reverse is usually true. Along the slopes of the adjacent mountains, some areas will be in the marine air during the day and in the dry, subsiding air at night. The relative humidity may begin to rise during the late afternoon and early evening and then suddenly drop to low values as dry air from aloft moves down the slopes. Abrupt humidity drops of up to 70 percent in the early evening have been observed.

SUMMARY

In this chapter we have considered atmospheric moisture in some detail. We have seen that moisture escapes into the atmosphere through evaporation from water bodies and soil, and through transpiration from vegetation. Atmospheric humidity is usually measured with a psychrometer and can be described in several ways. The dew-point temperature and the absolute humidity represent the actual moisture in the air, while the relative humidity indicates the degree of saturation at a given temperature.

We have also seen that absolute humidity varies in space and time for several reasons; however, relative humidity does not necessarily change in the same manner, because relative humidity is very dependent upon air temperature. The temperature effect frequently overrides the absolute humidity effect; therefore, relative humidity usually varies inversely with temperature.

While temperature and moisture distributions in the layer of air near the ground are important in fire weather because of their influence on fuel moisture, the distributions of temperature and moisture aloft can critically influence the behavior of wildland fire in other ways. The first of these influences will be seen in the next chapter when we consider atmospheric stability.