SOME ASPECTS OF FORMATION AND DISSIPATION OF FOG

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(Manuscript received March 3rd, 1939)

1. Introduction. The name *fog* is given to any cloud that envelops the observer and reduces the horizontal range of visibility to 1000 metres or less. If, under similar conditions, the horizontal range of visibility exceeds 1000 metres, the phenomenon is called *mist*[1]). This definition of fog is adequate at sea and over level ground, but it is inconsistent in hilly country, where the clouds may touch the hillsides without touching the lowland. In such cases, an observer in the lowland would report stratus while an observer at an elevated station would report fog. It is necessary to bear this distinction in mind both when analysing weather charts and when predicting fogs. What we are here concerned with is to discuss the formation of such fogs that may form in contact with the sea surface or level ground.

Non-saturated air may become saturated in three different ways, viz., (a) by evaporation of water into the air, (b) by mixing between horizontally or vertically adjacent air masses, and (c) by cooling. When fogs form, it is necessary that these processes should take place at the surface of the earth. In most cases of fog formation, the three processes operate together. We shall, however, first discuss each of them separately, and afterwards comment on their various combinations.

2. Evaporation. The evaporation of water, either from the underlying surface or from rain falling through the air, is proportional to the factor $(E - e)$, where E denotes the saturation vapour pressure corresponding to the temperature of the liquid water, and e is the actual water vapour pressure of the air, As e increases and approaches E, the evaporation decreases and approaches zero. It is convenient to consider three cases according to whether the air temperature is higher, equal to, or lower than the temperature of the liquid water. In the following, E_a denotes the saturation water vapour pressure of the air.

(a) *The temperature of the air is higher than that of the liquid water.* In this case, balance is reached when $E = e < E_a$, and the evaporation from the liquid water ceases before the air becomes saturated, because the saturation water vapour pressure of the air is higher than the saturation pressure corresponding to the temperature of the liquid water. If evaporation takes place from a terrestrial source of water, the heat of vaporization is supplied mostly by the water, and the air temperature remains sensibly unchanged. It follows then that such evaporation will not cause condensation in the air.

If evaporation occurs from falling rain, the heat of vaporization is supplied mostly by the air, which is then cooled. Through continued evaporation, the air temperature will approach the wet-bulb temperature, which remains constant during the process. The air is then saturated with water vapour ($E = e \gtreqless E_a$), and no further evaporation is possible from the falling rain. *As a fog consists not only of saturated air, but of air that contains a considerable amount of condensed water, it is evident that, when the temperature of the air is higher than that of the liquid water, a fog cannot form because of evaporation only.*

The conditions described above apply when the air is varmer than the underlying surface, and also when rain from colder air aloft falls through the varmer air below. The latter case is the normal occurrence in the atmosphere. Thus, when the air temperature decreases along the vertical, the falling rain-drops will be colder than the air through which they fall; the air will then remain non-saturated until it has been cooled to its wet-bulb

[1]) Definitions adopted by the International Meteorological Organization, 1929.

temperature, after which no further evaporation will occur.

(b) *The temperature of the air equals that of the liquid water.* In this case balance is reached when $E = e = E_a$; i. e. when the air has become saturated. As above, it is evident that evaporation alone will not cause condensation to occur in the air.

(c) *The temperature of the air is lower than that of the liquid water.* In this case water will evaporate into the air until $E = e > E_a$. If no condensation nuclei are present in the air, supersaturation will occur; but if such nuclei are present in sufficient amounts, the superfluous water will be condensed. In such cases, a fog, or a cloud, may result through condensation only.

Here, too, we may distinguish between (i) evaporation from falling rain, and (ii) evaporation from the underlying surface.

(i) When the air temperature increases along the vertical, (e. g. in inversion layers or along pronounced frontal surfaces), the falling rain-drops may be slightly warmer than the air through which they fall. If the air is not saturated, its wet-bulb temperature will be lower than the air temperature, and thus considerably lower than the temperature of the falling rain-drops. Through evaporation of the falling rain, the air is cooled to its wet-bulb temperature; condensation will then occur, and if the wet-bulb temperature of the air is sufficiently below the temperature of the falling rain-drops, condensation will occur on the condensation nuclei, and the droplets thus formed may grow in number and size to such an extent that the horizontal range of visibility is reduced 1000 metres or less. The cloud layer, from which the rain falls, then builds downwards and may eventually touch the earth's surface, and hence a fog results. Such fogs may form under pronounced frontal surfaces when the temperature of the air above the frontal surface is much higher than the temperature of the air below it (*frontal fogs*). The same applies when rain from a cloud system aloft falls through a layer of cold air under a ground inversion.

What has been said above is illustrated in Fig. 1. In all four diagrams the base of the rain-cloud is at *aa*. In the case *A*, the rain-drops will be colder than the air below *aa*. Fog or stratus will not then result from evaporation of falling rain. In the case *B*, the rain-drops will be warmer

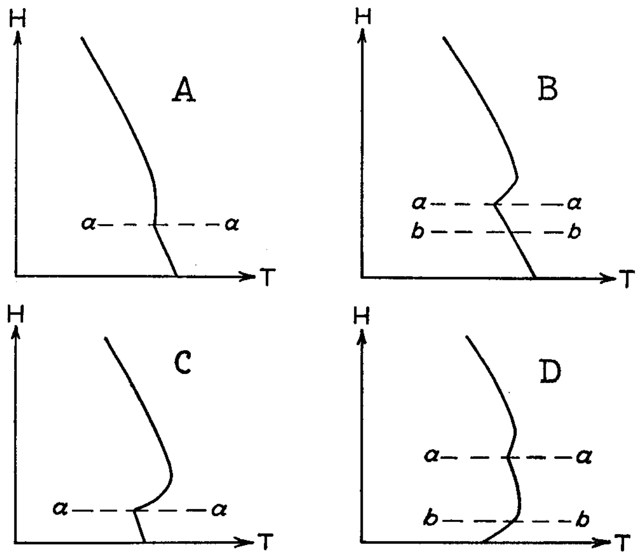

Fig. 1. Types of temperature-height curves in relation to fog or stratus under the frontal rain area.

than the air in the layer between *aa* and *bb*. A layer of stratus will then build down from the frontal cloud system; below the level *bb* the drops will be colder than the air, and condensation will not result. In the case *C*, evaporation from the falling rain may fill the whole layer under *aa* with fog. It is readily seen that such fogs can only develop when the frontal surface is close to the ground. In the case *D*, there is a shallow layer of cold air under a ground inversion. In this layer, the falling rain-drops will be warmer than the air; a fog may then result under *bb*.

It should be borne in mind that fogs formed by evaporation of rain from warm air aloft falling into cold air below can only occur when the temperature difference is sufficiently large, and when none of the fog-dissolving processes, which will be described later, overcompensate the effect of evaporation.

(ii) Fogs are sometimes observed when cold air streams over a water surface the temperature of which is very much higher than the air temperature. These fogs are known as *Steam Fogs*, or *Arctic Sea Smoke.* In such cases $E > E_a$, and the cause of the formation of such fogs is the intense evaporation from the water surface. For example, a water surface of 10° C would have a saturation vapour pressure of 12.3 mb. If the air above the water had a temperature of 5° C and a relative humidity of 100 percent, the vapour pressure of the air would be 8.7 mb., and the difference in water vapour pressure between the surface and the air would be 3.6 mb. Water would evaporate quickly and fill the air with vapour: steam would pour forth from the surface, and the air be filled with a steam fog.

It is important to bear in mind that extreme temperature differences are required to produce a steam fog. This condition is due to the circumstance that the cold air above a warm surface will be heated from below to such an extent that the air becomes unstable. Through instability, vertical mixing sets in and prevents the steam from accumulating in the air. In general, light winds and a strong pre-existing inversion near the sea surface are necessary for the steam to accumulate in the surface layer. In extreme cases, steam fogs are known to occur in strong winds. Thus, during a cold spell in February of 1934, steam fogs occured at the harbour of East Boston[1]); the air temperature was then about — 28° C, probably 30° C lower than the sea surface temperature.

Steam fogs are most frequently observed along the arctic coasts and in the fjords of cold continents (e. g. Norway, where they are called «frostrøk» (frost-smoke)); they may also occur in the autumn in cold continental air masses over lakes and rivers.

3. Horizontal Mixing. The wind is never a steady current: it consists of a succession of gusts and lulls of short period. This condition is mainly due to eddies that form on account of friction at the ground, horizontal and vertical stresses, and instability. An eddy that moves from one place to another may be regarded as an agency in the process of diffusion of heat, moisture, momentum, etc.

The mixing that occurs between horizontally adjacent air masses may be assumed to take place at constant pressure. Let M_1, T_1, e_1 and q_1 represent the total mass, the absolute temperature, the vapour pressure, and the specific humidity in the first component of the mixture; and let similar variables with the subscript 2 represent the second mass. Let T, e, and q represent the temperature, the vapour pressure, and the specific humidity in the air after complete mixing at constant pressure. The following formulae then hold:

$$q = \frac{M_1 q_1 + M_2 q_2}{M_1 + M_2} \tag{1}$$

$$e = \frac{M_1 e_1 + M_2 e_2}{M_1 + M_2} \tag{2}$$

$$T = \frac{M_1 T_1 + M_2 T_2}{M_1 + M_2} \tag{3}$$

The last formula is not exact, but is holds with sufficient accuracy for all practical purpose[1]).

It is readily seen that a formula similar to (3) holds for the equivalent temperature, but not for the wet-bulb temperature. However, on account of the relation which exists between the equivalent temperature and the wet-bulb temperature, the latter is readily obtained from the former, e. g. by aid of a thermodynamical diagram.

Let the points A and B in Fig. 2 represent two parcels of air. As both points fall to the right of the saturation line, it follows that neither are saturated. If the two parcels were completely mixed into each other, the conditions of the mixture would be represented by a point C on the straight line AB, C indicating the common centre of gravity of the two parcels. As the saturation water vapour pressure is not a linear function of the air temperature, it follows that the mixture of two non-saturated parcels may be saturated, or super-saturated. In the latter case, the superfluous water will be condensed as a fog or a cloud. From this it has been concluded that mixing is a frequent cause of formation of clouds or fogs. This conclusion, however, requires a more definite proof in so far as the orders of magnitudes of the temperature and humidity differences between adjacent air masses are concerned; and it is also necessary to consider the effect of vertical mixing.

In order to show more accurately the maximum amounts of superfluous water which may result from complete mixing, we assume that $M_1 = M_2$, so that $T = \frac{1}{2}(T_1 + T_2)$, and $e = \frac{1}{2}(e_1 + e_2)$.

We let subscript 1 indicate the warmer mass, and R_1 denote the relative humidity corresponding to T_1 and e_1; and R_2 the relative humidity corresponding to T_2 and e_2. Table 1 then shows the resulting relative humidity, or, in the case of supersaturation, the amount of superfluous water expressed in grammes per cubic metre.

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[1]) *Byers*, H. R. Synoptic and Aeronautical Meteorology. New-York, 1937 p. 192.

[1]) *Brunt*, D. Physical and Dynamical Meteorology. Cambridge, 1934.

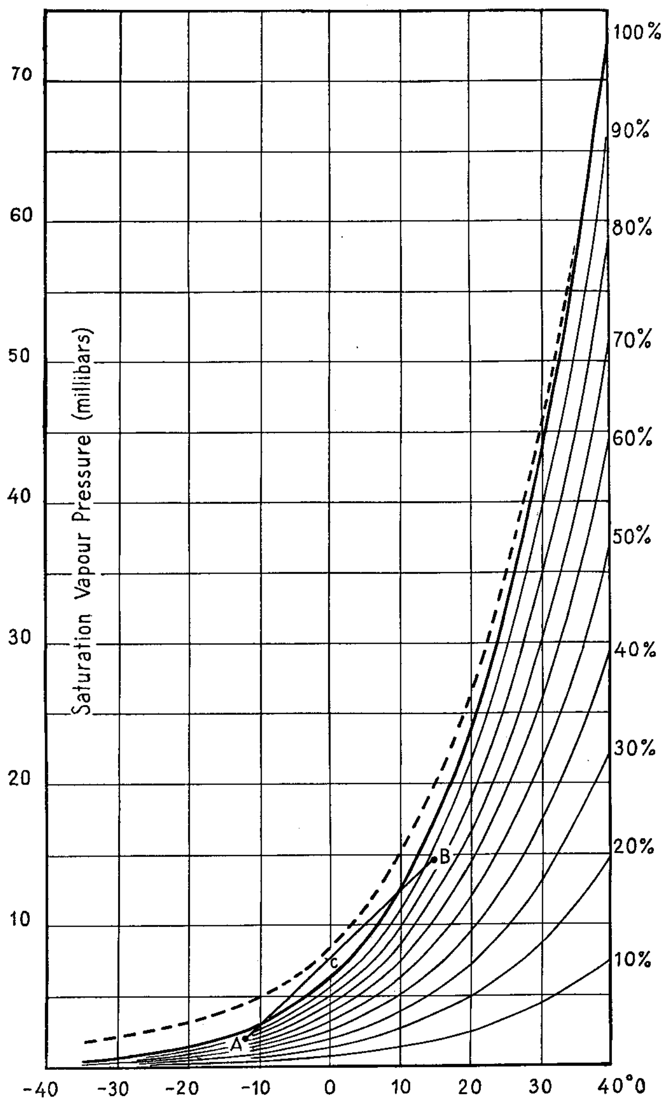

Fig. 2. Relation between the water vapour pressure and air temperature for standard values of relative humidity. The heavy curve represents saturation ($R = 100$ percent). The broken curve indicates the super-saturation which corresponds to saturated air plus 0.5 grammes liquid water per cubic metre of air.

It is known that the content of liquid water in clouds and fogs may vary from 0.1 to 5.0 grammes per cubic metre. The horizontal distance between the saturation curve and the broken curve in Fig. 2 represents the cooling which saturated air must undergo in order to cause so much super-saturation that the superfluous water amounts to 0.5 grammes per cubic metre. If this water were condensed, a fog or a cloud of slight density would result. It is readily seen from Table 1 that the temperature differences which normally occur between adjacent air masses in nature are altogether insufficient for horizontal mixing to produce so much condensation that 0.5 grammes of liquid water per cubic metre is liberated.

It will be seen from Table 1 and from Fig. 2, that mixing at high temperatures will result in condensation only when both components are close to saturation. This condition is due to the small curvature of the saturation curve when the air temperature is high. Furthermore, other conditions being equal, the amount of superfluous water increases with increasing temperature (T), because of the mutual convergence of the saturation curve and the super-saturation curve. The conditions for super-saturation to occur are more favourable when the warmer component has the higher relative humidity. At low temperature, the colder component may be relatively dry; this condition is due to the convergence with decreasing temperatures of the relative humidity curves in Fig. 2.

The table shows that the amount of superfluous water is very small even under the most favourable conditions. It is important to note that the table has been computed on the assumption that there is a sharp discontinuity between the two components. If there were a gradual transition between them, the amount of superfluous water would be considerably reduced. Moreover, when condensation occurs, the latent heat is liberated, and the air temperature would be raised accordingly; this counteracts the cooling by mixing and the amount of superfluous water. The point which represents the final result of mixing would, when the liberated latent heat is accounted for, fall to the right of and below the point C in Fig. 2. The actual amount of superfluous water may be computed by means of the Clausius-Clapeyron equation[1]). It suffices here to remark that the amounts of superfluous water shown in Table 1 are greatly in excess of that actually occurring in the atmosphere. It follows then that *horizontal mixing cannot cause so much condensation that a fog or a cloud results.* Under favourable conditions, it may, however, produce a mist. In the above arguments, we have assumed that the mixing takes place entirely in the horizontal direction. Conditions are widely different when the mixing occurs mainly along the vertical.

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[1]) *Brunt*, D. Condensation by Mixing. Q. J. of Royal Met. Soc. Vol. *61*, 1935.

Table 1.

Relative Humidity or Amount of Superfluous Water Resulting from Complete Horizontal Mixing.

$T_1 - T_2$	R_1 \ R_2	$T = -20°$ C 100 %	95 %	90 %	$T = 0°$ C 100 %	95 %	90 %	$T = 20°$ C 100 %	95 %	90 %
2°	100 %	*0.00*	98.1	95.7	*0.01*	98.0	95.6	*0.03*	97.8	95.5
	95 %	97.5	95.3	93.0	97.6	95.3	92.9	97.5	95.2	92.8
	90 %	94.7	92.6	90.3	94.9	92.6	90.3	95.3	92.5	90.1
4°	100 %	*0.01*	99.3	97.2	*0.05*	98.8	96.6	*0.12*	98.5	96.3
	95 %	98.4	96.3	94.2	98.1	95.9	93.8	97.8	95.6	93.4
	90 %	95.4	93.3	91.2	95.2	93.0	90.9	95.0	92.8	90.6
6°	100 %	*0.03*	*0.01*	99.2	*0.10*	*0.01*	98.1	*0.26*	99.9	97.8
	95 %	99.8	97.9	96.0	99.0	97.0	95.0	98.5	96.9	94.8
	90 %	96.6	94.7	92.8	95.9	93.9	91.9	95.5	93.9	91.8
8°	100 %	*0.06*	*0.04*	*0.02*	*0.18*	*0.09*	*0.00*	*0.46*	*0.13*	98.8
	95 %	*0.02*	*0.00*	98.4	*0.02*	98.6	96.8	99.5	97.5	95.5
	90 %	98.4	96.4	94.9	97.1	95.3	93.4	96.3	94.4	92.4

4. Vertical Mixing. In this case the pressure is not constant on the individual air particles, and it is necessary to consider adiabatic changes. The eddy flux of heat in non-saturated air across a horizontal unit area per unit time is expressed by[1])

$$F_1 = -K_1 c_p \varrho (\gamma_d - \gamma)$$

Here K_1 is the coefficient of eddy transfer of heat, c_p is the specific heat of air at constant pressure, ϱ is the density of the air, γ_d and γ denote the dry-adiabatic lapse-rate and the actual lapse-rate of the air, respectively. If the air were saturated, γ_m (the moist-adiabatic lapse-rate) should be substituted for γ_d. If no heat is supplied to, or withdrawn from, the air from some external source, and if no condensation takes place, the mean temperature of the air column will remain constant throughout the process. Heat would then be transported downwards when $\gamma_d > \gamma$; when $\gamma > \gamma_d$, heat would be transported upwards by the eddies. The flux of heat along the vertical would cease when $\gamma = \gamma_d$. The final result of vertical mixing would be characterized by:

$$(1) \qquad \gamma = \gamma_d, \text{ or } \frac{\partial \theta}{\partial z} = 0,$$

and the mean temperature and the mean potential temperature (θ) of the air column unchanged.

In a similar manner, the eddy flux of specific humidity (q) along the vertical is expressed by:

$$F_2 = -K_2 \frac{\partial q}{\partial z},$$

where K_2 is the coefficient of eddy transfer of specific humidity. It will be seen that moisture is transported upwards when $\frac{\partial q}{\partial z} < 0$, and downwards when $\frac{\partial q}{\partial z} > 0$. If no moisture is supplied to, or withdrawn from, the air, and if condensation does not occur, the mean specific humidity of the air column will remain unchanged throughout the process. The final result of vertical mixing will then be characterized by:

$$(2) \qquad \frac{\partial q}{\partial z} = 0,$$

where q = the mean specific humidity of the air column at the initial moment.

It is readily seen (e. g. from a thermodynamical diagram) that the equations (1) and (2) imply

$$(3) \qquad \frac{\partial T_w}{\partial z} = \gamma_m \quad \text{or} \quad \frac{\partial \theta_w}{\partial z} = 0,$$

where T_w denotes the wet-bulb temperature, and θ_w denotes the wet-bulb potential temperature of the air after complete mixing.

[1]) See *Brunt*, D., Physical and Dynamical Meteorology, Cambridge, 1934.

Under normal conditions, the air is stably stratified and the specific humidity decreases along the vertical. Vertical mixing would then transport heat downwards from the top of the mixed layer, and moisture upwards to the air that is cooled by mixing. It is readily seen that this may result in the condensation of water vapour. It is, however, important to note that vertical mixing can cause condensation only in the upper portion of the mixed layer. The lowest level at which condensation occurs as a result of vertical mixing is called *the mixing condensation level.* How this level is determined by means of aerological ascents has been shown by *Petterssen*[1]).

Fig. 3 shows, as an example, the result of vertical mixing in a limited layer of air which is

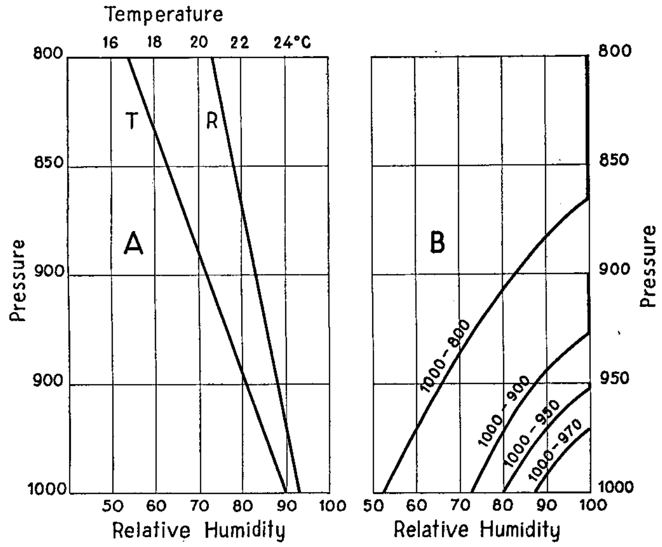

Fig. 3 A. The initial state. T = air temperature; R = relative humidity.

Fig. 3 B. Vertical distribution of relative humidity after complete stirring. The uppermost curve indicates the distribution of relative humidity when the entire layer between 1000 and 800 mb. is completely stirred; the next curve shows the result when the layer between 1000 and 900 mb. is stirred, etc.

originally stably stratified and non-saturated. It will be seen that the mixing condensation level ($R = 100$ percent) varies with the thickness of the stirred layer. Thus, when the layer between 1000 and 950 mb. is completely stirred, condensation occurs only in a very thin layer. If the mixed layer is shallower than that indicated by the lower right curve (1000 — 970 mb.), condensation does not result.

[1]) Weather Analysis and Forecasting. Not yet published.

That vertical mixing of non-saturated air does not cause condensation at the base of the mixed layer is readily verified by plotting the initial values of T and q on a thermodynamical diagram, and applying the conditions which characterize the final state. It will then be seen that the conditions for condensation to occur are most favourable when the initial state is characterized by $\frac{\partial T}{\partial z} > 0$ and $\frac{\partial q}{\partial z} > 0$; but, even then, the air at the base of the layer remains non-saturated for such inversions of temperature and humidity which occur in the atmosphere.

From what has been said above it follows that vertical mixing is a potent factor in the formation of stratus. This is particularly true when the mixed layer is limited by a pre-existing inversion. On the other hand, as vertical mixing always causes the air in the lower portion of the stirred layer to remain non-saturated, a fog cannot form as a result of vertical mixing. In fact, vertical mixing is highly effective in dissipating existing fogs and strongly counteracts the fog producing agencies.

It was shown in para. 3 that horizontal mixing could only cause insignificant amounts of condensed water in the air. As horizontal mixing is always associated with vertical mixing (and vice versa), it follows that the former is counteracted and overcompensated by the latter.

If mixing (vertical and horizontal) occurs together with rain falling from warmer air aloft (see para. 2), the stratus layer caused by mixing and intensified by evaporation, may build downwards, and may eventually reach down to the sea surface or level ground, thus forming a real fog. The conditions for this to occur are particularly favourable in the air under frontal rain areas as shown in Fig. 1 C. Such fogs are therefore often called frontal fogs even though they form mainly as a result of evaporation and mixing. Again, it should be emphasized that evaporation plays an important rôle in the formation of such fogs only when the air from which the rain falls is considerably warmer than the air near the earth. This happens quite frequently in cold continents in winter. The conditions are then as shown in Fig. 1 D.

It is, of course, obvious that the formation of such fogs is greatly facilitated when the surface air is simultaneously cooled by the underlying surface.

5. Cooling. By far the most frequent and the most effective cause of fog formation is cooling of the air while it is in contact with the underlying surface. This cooling may be due to several causes. From the first theorem of thermodynamics we obtain:

$$(1) \qquad \frac{dT}{dt} = \frac{1}{c_p}\frac{dQ}{dt} + \frac{R}{c_p}\frac{T}{p}\frac{dp}{dt},$$

where $\frac{dQ}{dt}$ denotes the heat added to, or withdrawn from, a unit mass of air per unit time. Let $\boldsymbol{v}$ denote the horizontal wind vector, w the vertical component of the wind velocity, and ∇p the horizontal pressure gradient. The above equation may be written in the form:

$$(2) \qquad \frac{dT}{dt} = \frac{1}{c_p}\frac{dQ}{dt} + \frac{RT}{c_p p}\frac{\partial p}{\partial t} + \frac{RT}{c_p p}\boldsymbol{v}\cdot\nabla p + \frac{RT}{c_p p}w\frac{\partial p}{\partial z}$$

As we are here concerned with the formation and dissolution of fogs, we need only consider motion of the air along the earth's surface. Eq. (2) then shows that there are four factors which control the temperature changes of the moving air, namely: (a) the non-adiabatic heating or cooling represented by the term containing $\frac{dQ}{dt}$, and three adiabatic influences resulting from (b) the local pressure variation $\left(\frac{\partial p}{\partial t}\right)$, (c) the flux of air across the isobars $(\boldsymbol{v}\cdot\nabla p)$, and (d) vertical motion $\left(w\frac{\partial p}{\partial z}\right)$. It is convenient to discuss the various terms in the opposite order.

(1) *The slope effect.* The term $\frac{RT}{c_p p}w\frac{\partial p}{\partial z}$ vanishes at sea level and over level ground, where $w = 0$; but when the wind blows up or down a slope, the term may become larger than any of the other terms in eq. (2). Even with a considerable up-slope velocity, the air produces a fog only when it is stably stratified in the saturated state, because, if it were not, convective currents would develop. This implies that the air must also be *convectively stable* $\left(\text{i. e. } \frac{\partial \theta_w}{\partial z} > 0\right)$ before saturation occurs, if an up-slope fog is to develop. Furthermore, the relative humidity must be high enough for the air to be cooled to its dew-point during the ascent; and the vertical mixing must be slight or moderate. Thus, *convective stability, slight turbulence, and high relative humidity are favourable conditions for the production of up-slope fogs.*

It is typical of up-slope fogs that they are very deep; this condition is due to the circumstance that the up-slope velocity affects a deep layer of air. Up-slope fogs are frequently observed with easterly winds in southeast Norway, and also on the eastern slope of the Rocky Mountains. The «Cheyenne Type» of fog belongs to this category.

(2) *The flux across the isobars.* The term $\frac{RT}{c_p p}\boldsymbol{v}\cdot\nabla p$ is directly proportional to the pressure gradient and the wind component normal to the isobars. This term can be important only when the pressure gradient is strong. But, when this is the case, the wind velocity will be high, and the vertical mixing will be appreciable and will overcompensate the effect arising from the flux of air across the isobars. As an example we assume that the horizontal pressure gradient is 2.10^{-6} centibar per metre (which is a high value), and that the wind component normal to the isobars is 5 m.sec.$^{-1}$. At normal pressure and temperature, the term under discussion would only amount to 0.03° C per hour, or 1° C in 33 hours. The term is therefore quite insignificant in comparison with the other effects, and need not be considered.

(3) *Local pressure changes.* The term $\frac{RT}{c_p p}\frac{\partial p}{\partial t}$ is important only when the air streams for a lengthy interval of time in an area of rapidly falling pressure. Let us assume that the barometer is falling 1 mb. per hour. The above term would then amount to 0.08° C per hour, or about 1° C in 13 hours.

Barometric tendencies larger than these occur quite frequently in high and middle latitudes, but they are then accompanied by strong winds (strong pressure gradients). In such cases vertical mixing is usually intense and counteracts the cooling influence caused by falling pressure. It should also be noted that the cooling caused by falling pressure is not confined to the layer of air close to the earth's surfac, because $\frac{\partial p}{\partial t}$ is normally uniform throughout a deep layer of air. As p decreases more rapidly with elevation than does T, it follows that the term under discussion is apt to exert a greater influence in the upper atmosphere than at the earth's surface, where, in addition, vertical mixing is most intense.

The conditions favourable for a fog to form on account of falling pressure are: (1) an intense isallobaric system which moves in the direction of the air current, (2) high relative humidity, (3) moderate or slight wind velocity, and, (4) stable stratification so as to prevent vertical mixing. There is no evidence in support of the view that a fog forms only as a result of falling pressure, but when the above conditions are favourable, the influence of falling pressure may add to those of the other fog producing agencies.

(4) *Non-adiabatic cooling.* The term $\frac{1}{c_p}\frac{dQ}{dt}$ constitutes the principal factor in the production of fogs. It is customary to distinguish between two principal types of fog formation, viz., *radiation fog* and *advection fog*, the former being principally caused by raditional cooling of the underlying surface; and the latter being mainly due to advection of warmer air over a colder surface. In both cases the air gives off heat to the underlying surface. Most frequently fogs form on account of the combined influence of radiative and advective cooling.

(i) *Radiation.* The influence of the radiative cooling over land on the diurnal variation in air temperature is usually of the order of magnitude of 1° C per hour. It is normally smaller than the up-slope effect discussed above, and it is many times larger than the adiabatic cooling caused by falling pressure and flux of air across the isobars. The radiative influence is of a periodical nature, and, strictly speaking, if no other process were in operation, it would either produce fog every night, or no fog at all. In discussing radiation fog it is necessary to consider also the motion of the air.

We consider first the case when stably stratified, non-saturated air streams along the isotherms of the underlying surface, so that there is no advective cooling. The eddy flux of heat (see para. 4) is then directed downwards, and this counteracts the radiative cooling of the underlying surface. Under normal atmospheric conditions, the specific humidity decreases upwards, and the eddies transport water vapour upwards through the air. Thus, *stable stratification and decreasing specific humidity along the vertical together with turbulence are unfavourable for formation of radiation fog.*

However, when the wind velocity decreases, as it normally does in the evening, the turbulence decreases rapidly (see para. 6), and so also do the downward flux of heat and the upward flux of moisture. If the relative humidity is sufficiently high, outgoing radiation from the surface may cool the surface air below its dew-point, and a fog may form. As long as the specific humidity decreases along the vertical, fog does not usually form unless in still air, and even then the cooling may only result in dew or rime on the ground.

When the specific humidity increases along the vertical, the eddy flux of moisture is towards the surface, and the conditions are favourable for formation of radiation fogs if the relative humidity is high, and the wind is so slight that the radiative cooling of the surface is not substantially reduced by the eddy flux of heat from above. In addition, the sky must be clear or lightly clouded, because otherwise the counter-radiation from the sky would reduce the outgoing radiation from the surface.

Suming up, we may say that the conditions favourable for formation of radiation fogs are: (a) *high relative humidity*, (b) *cloudless sky*, (c) *constant or increasing specific humidity along the vertical*, and (d) *lack of turbulence*. It should be noted that stable stratification involves a flux of heat towards the underlying surface which is unfavourable for radiative cooling. On the other hand, as the turbulence depends greatly on the stability of the air, point (d) above may be replaced by (e) *stable stratification and lack of wind.*

The above conditions (b) — (d) occur most frequently in continental anticyclones in the colder part of the year, or in any uniform pressure distribution when there is enough subsidence in the upper air to dissolve the clouds. As high relative humidity is an important factor, radiation fogs develop most frequently in air of maritime origin when it becomes stagnant over cold continents.

On account of the slight diurnal variation of temperature at sea, radiation fogs do not easily develop over oceans.

(ii) *Advection.* The term $\frac{1}{c_p}\frac{dQ}{dt}$ is an important factor when warmer air streams over a colder surface. The cooling of the travelling air depends on the difference in temperature between the air and the underlying surface, which again depends on the speed with which the air streams across the isotherms of the surface. In extreme cases the cooling may amount to 2° C per hour, but it is usually less than 1° C per hour. As high wind velocities (strong turbulence and vertical mixing) are highly unfavourable for the formation of fog, and as a certain wind velocity is necessary to bring the air across the isotherms of the underlying surface, it follows that advection fogs develop as a compromise between the effects of vertical mixing and cooling from below. This explains why fogs are comparatively rare in air currents moving from warmer towards colder regions. The conditions favourable for the formation of advection fogs are: (a) *large difference between the temperature of the air and that of the underlying surface,* (b) *not too high wind velocity,* (c) *high relative humidity initially,* and (d) *stable stratification initially.*

The factors (a) and (b) are of greatest importance. As turbulent mixing is roughly proportional to the wind velocity, and as large temperature difference can be maintained only when the air streams across the isotherms of the underlying surface, it follows that the best condition for advection fog to form is when the air streams with a moderate velocity normal to the isotherms of the underlying surface, because the advective cooling will then be greatest in comparison with the downward eddy flux of heat.

Practically all sea fogs are of the advection type. On land, radiation fogs are predominant. These usually form in stagnant air from a warm and humid source which subsequently becomes cooled by outgoing radiation from the earth's surface. In such cases the cause of the fog formation may be ascribed to *advection followed by radiation.*

6. Fog in Relation to Wind. From what has been said above it follows that turbulence is a factor of basic importance, inasmuch as the vertical mixing which results from turbulence is highly effective in counteracting the fog producing agencies. The intensity of turbulence depends mainly on two factors: the stability of the air and the velocity of the wind. That this is so is shown in a detailed study of fog by *Taylor*[1]) from which the following passage is quoted (refer to Fig. 4):

«It will be noticed that on all these days the wind speed was about 10 miles per hour both during the day and during the night. On the three days 5th–6th, 8th–9th, and 9th–10th the wind was gusty during the day, and practically free from gusts during the night. The anemometer trace for July 6–7, however, shows gustiness throughout the whole 24 hours, though the amount of gustiness at night is rather less than it is during the day.

The reason for this difference may be inferred from the record of the thermograph for the week July 3–10, which is also shown in Fig. 4. It will be seen that, during the evenings and nights of the 5th, 8th, and 9th the temperature fell through 26°, 25°, and 17° F. On the 6th, however, the temperature stopped falling at about 8.30 p.m. (owing to a sheet of cloud which came over the sky) after it had fallen only 9° F below the maximum during the day. It seems, therefore, that the temperature gradient, due to the cooling of the ground on a clear night when the temperature of the surface air falls about 20° F, is sufficient to suppress practically all turbulent motion in a wind of 10 m.p.h. at a height of 140 feet (the height of Pyestock chimney), while the temperature gradient due to cooling of 9° F on a cloudy night is not sufficient to suppress the turbulent motion due to the friction of a wind of about 10 m.p.h. over the ground.»

The intensity of turbulence increases with the speed of the wind; and when the wind increases beyond a certain value (which depends on the stability of the air) the fog may be dissipated or develop into a layer of stratus. The effect of turbulent mixing accompanying strong winds is shown clearly in the following table which is quoted from *Taylor*'s paper.

Table 2.

Wind Force Beaufort scale	Number of cases of fog	Mean value of air temperature minus water temperature
0	3	−0.5
1	20	0.3
2	30	1.0
3	46	1.0
4	26	1.1
5	12	1.8
6	3	2.6
7	1	4.1

It will be noted that the maximum frequency is found neither at the extremely high values of the difference in temperature between air and sea, nor at the lowest wind of velocities, but somewhere between. This condition is due to the following circumstances: Great temperature difference occur only when the air moves rapidly across

[1]) *Taylor,* G. I. The Formation of Fog and Mist. Quart. Jour. Roy. Met. Soc., vol. 43, pp. 241–268, 1917.

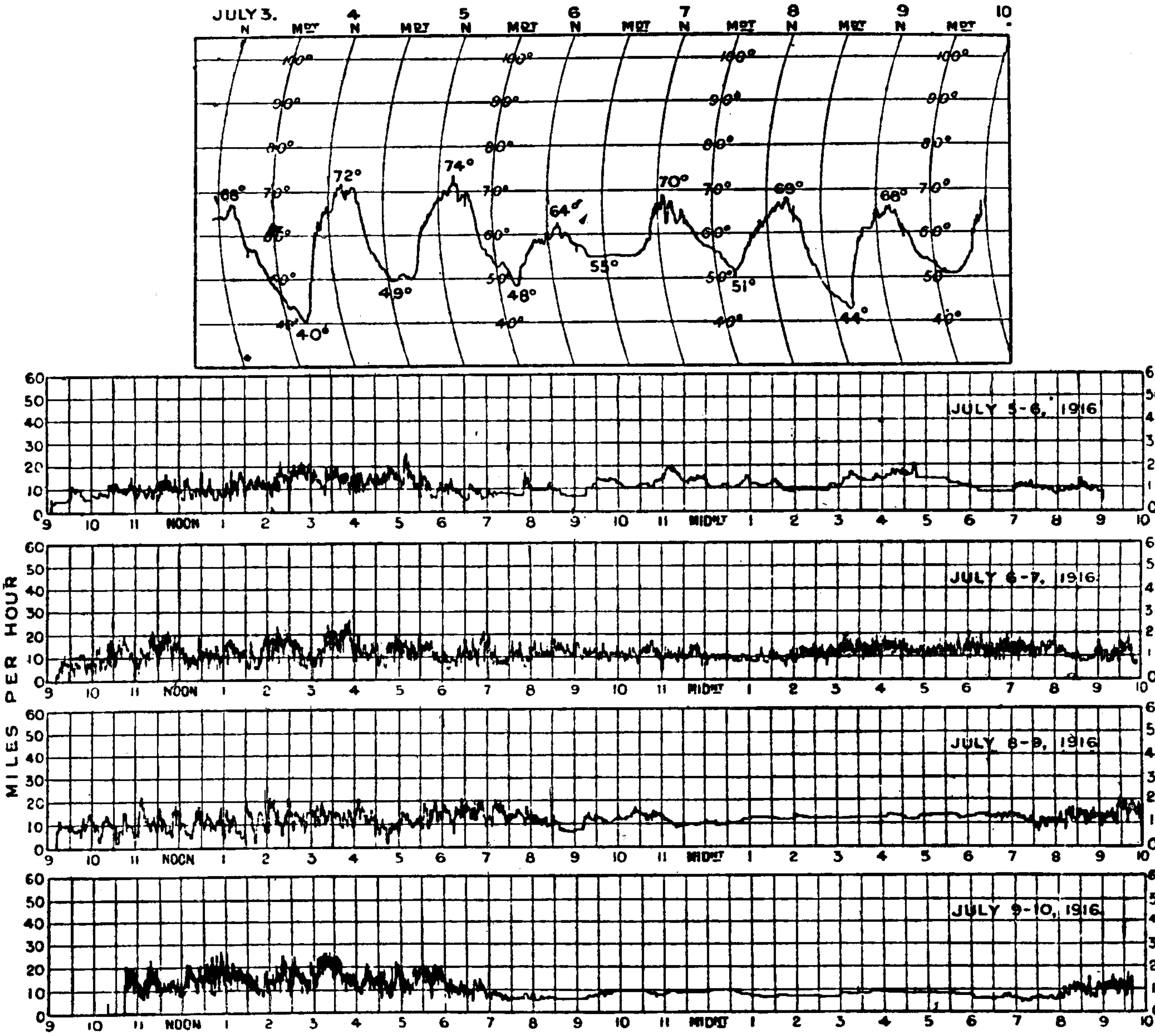

Fig. 4. Anemograms from Pyestock, near Farnborough, 140 feet above the ground, and the corresponding thermogram.

the isotherms of the surface; but in such cases fogs are rare on account of intense mixing. On the other hand, when the wind velocity is slight, the difference in temperature is also slight; the air is then less stable, and even slight turbulence suffices to dissipate the fog. Thus, sea fogs (which are advection fogs) are most likely to form when the wind speed is moderate and when it is blowing more or less directly across the isotherms of the water surface. The above table shows that fogs may sometimes persist in air that is slightly colder than the sea surface. This only happens in almost still air. Such fogs must have formed while the air was blowing from warm towards colder sea[1]).

Table 2 refers to advection fog. Over land it is difficult to distinguish between advection fog and radiation fog. Table 3, which is quoted from *Taylor*'s paper, shows the frequencies of wind forces at different hours on the 70 occasions when fog was reported at Kew in the nights during the years 1900—1905.

[1]) «Steam fog» could only form if the air temperature were considerably below the sea surface temperature (see para. 2).

Table 3.
Frequencies of Wind Forces Preceding Night-Fogs at Kew, 1900—1905.

Wind Velocity m.p.h.	Wind Force Beaufort scale	4 p.m.	6 p.m.	8 p.m.	10 p.m.	Mid-night
0– 3.3	0–1	24	35	50	58	62
3.3– 5.5	1–2	23	20	18	10	5
5.5– 9.2	2–3	16	13	1	2	3
9.2–13.6	3–4	7	2	1	0	0

The table shows the normal decrease in wind velocity from the afternoon towards midnight. On looking at the 6 p.m. column of Table 3, it will be seen that only twice in five years was a wind greater than 9.2 m.p.h. at 6 p.m. followed by a fog in the night. Similarly, only twice in five years was a wind greater than 5.5 m.p.h. at 8 p.m. followed by fog in the night. The table may be used for forecasting, because if «no fog» had been predicted at Kew on every occasion when the wind at 8 p.m. was greater than 5.5 m.p.h., only two mistakes would have been made in five years.

Thus, while it is practically certain that a fog does not form at Kew when the wind at 8 p.m. is greater than 5.5 m.p.h., there, naturally, remains a great number of cases when the wind at 8 p.m. is less than 5.5 m.p.h. without fog forming later in the night. The reason for this is that the formation of fog also depends on the temperature of the surface and the temperature and the humidity of the air.

It is of interest to remark that while Table 3 holds only for Kew, there can be little doubt that it is applicable, in a general way, also to other land stations similarly situated. In view of the many local factors which influence fog formation on land, it is highly probable that local statistical studies, on the lines indicated by *Taylor*, would substantially improve the forecasting of fog.

On comparing Table 2 with Table 3, it will be seen that the upper limit of wind force at which fog can exist is much greater at sea than at Kew. This is partly due to the increased friction over land which causes increased vertical mixing; and partly to the circumstance that the land fogs are mostly radiation fogs, which cannot form unless the wind is slight, while the sea fogs are advection fogs, which require a certain wind velocity to bring the air across the isotherms of the underlying surface.

7. Fog in Relation to Temperature. Whether the fog is produced by radiation, advection, or any of the adiabatic processes described in para. 5, the forecaster will have to estimate the amount of cooling that is likely to occur within the forecasting period. Whether the expected cooling suffices to produce a fog or not, depends on the humidity of the air.

Fig. 5 shows a diagram from which the amount of cooling necessary to produce a fog can be determined when the air temperature and the relative humidity are known. Let A denote a point in the diagram to which correspond a certain relative humidity and a certain air temperature. The horizontal lines then indicate the corresponding dew-point. The distance from A to B measures the amount of cooling necessary to make the air saturated, provided that the moisture content remains constant during the cooling.

It should be borne in mind that fog is not merely saturated air, but air which contains an amount of liquid water sufficient to reduce the horizontal range of visibility to 1000 metres or less. The visibility is primarily a function of the amount of liquid water, but it also depends on the size of the droplets and the pollution. As it is known that the amount of liquid water in fogs and clouds may vary between 0.1 and 5.0 grammes per cubic metre, it seems probable that a moderate fog would not contain less than 0.5 grammes liquid water per cubic metre. Whether this value is exact or not is a matter of minor importance. What we wish to emphasize here is that after the air has become saturated, a certain amount of cooling is necessary in order to produce so much condensed water that the horizontal visibility is reduced to 1000 metres or less. This is shown in Fig. 5, where, for example, the air represented by the point A must be cooled by the amount AB in order for it to become saturated, and thereafter it must be cooled still further, by the amount $\Delta T = BC$, in order that it should contain 0.5 grammes of liquid water per cubic metre.

While the air is being cooled from B to C, it will be misty, but a real fog would not develop until the amount of liquid water has reached a certain value. It is of interest to note that the

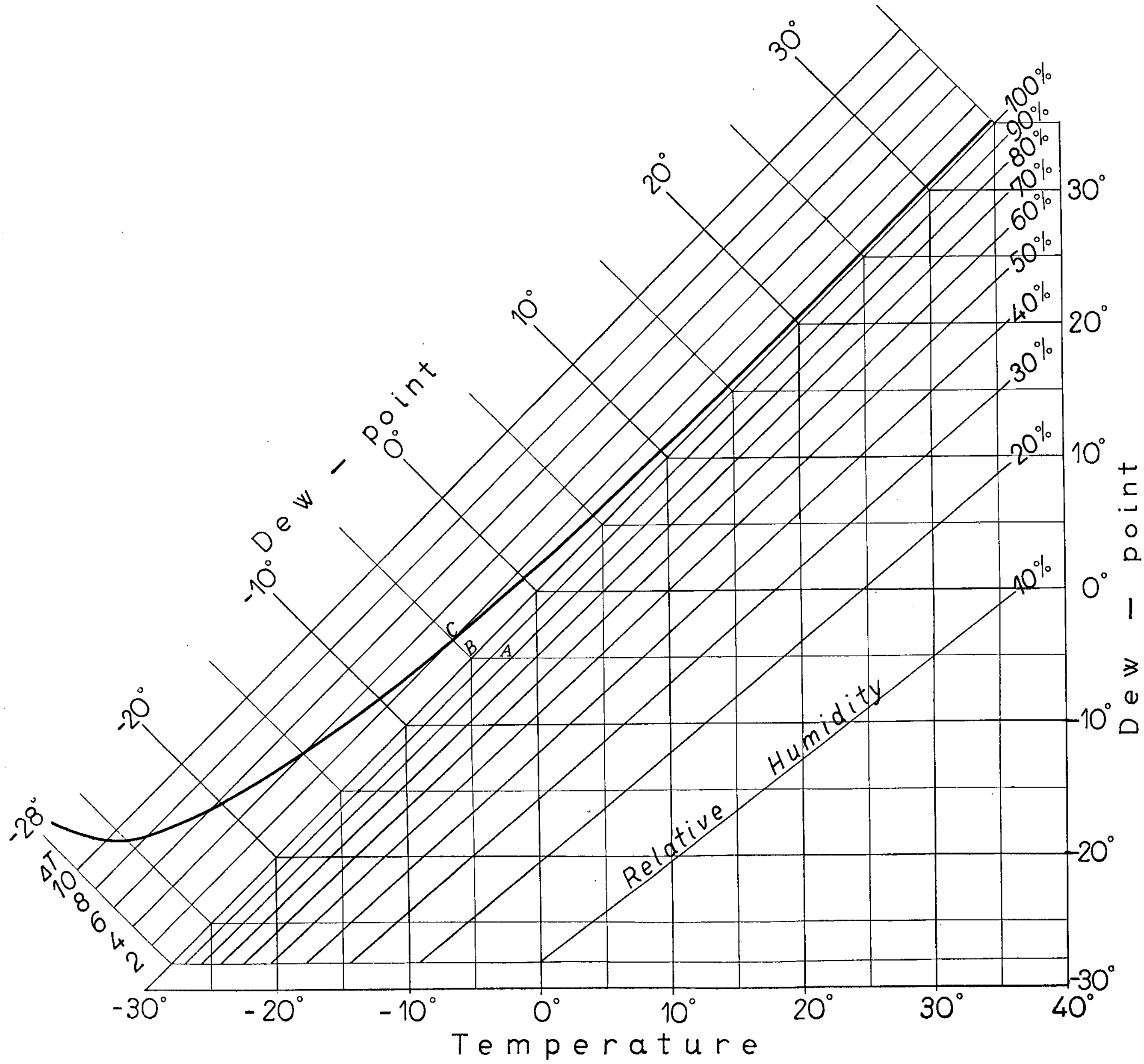

Fig. 5. Fog prediction diagram.

«*mist interval*» is very small when the air temperature is high, and it increases as the temperature decreases. When the temperature approaches — 28° C, the «mist interval» widens rapidly, because almost the total amount of aqueous vapour in the air must then be condensed in order to produce a sufficient amount of liquid water. Thus, at — 25° C, saturated air must be cooled by about 9° C for 0.5 gr. water to be condensed, whereas at + 20° C, a cooling of 0.4° C produces the same effect. It follows then that at low temperatures, the amount of cooling necessary to produce a fog becomes greater than the diurnal amplitude of temperature, and no fog will form solely as a result of nocturnal cooling. We shall see later that the drying influence of the snow has a tendency to counteract the formation of fog over snow-covered ground, and to dissipate fog that is transported to such areas.

For prediction of radiation fogs it is of basic importance to estimate the amount of cooling which is likely to occur. As radiation fogs do not form when the wind velocity is high, or when the sky is cloudy, it suffices to consider the conditions when the cloudiness is small and when the wind velocity is slight or moderate (see para. 6). Suppose that the observation is made at a certain (h), and that we know the normal diurnal fall in temperature under calm and cloudless conditions after the hour h. From the temperature-moisture diagram (Fig. 5) we may determine the amount of cooling necessary to produce saturation. If this amount is less than the normal fall in temperature, fog is likely to develop; and if it is greater, fog is unlikely.

It is well known that the diurnal amplitude of temperature for any given locality is greater on warm days (in the warm season) than on cold days (in the cold season). The normal fall in temperature on almost calm and clear nights must be determined statistically for representative stations. It is then possible to draw a line on a temperature-humidity diagram in such a way that it separates the combinations of temperature and humidity at which fog can form, from those at which fog is improbable. Fig. 6 shows such a diagram prepared by *Taylor* (loc. cit.) for Kew. *Taylor* found that the same diagram holds also for Oxford, Nottingham and Potsdam. As regards the method of construction of such diagrams, the reader is referred to *Taylor*'s paper. This method of fog prediction is particularly suitable for short range forecasts, such as are required at aerodromes.

In estimating the chances for formation of radiation fogs it is necessary to know not only the normal variation of air temperature on clear,

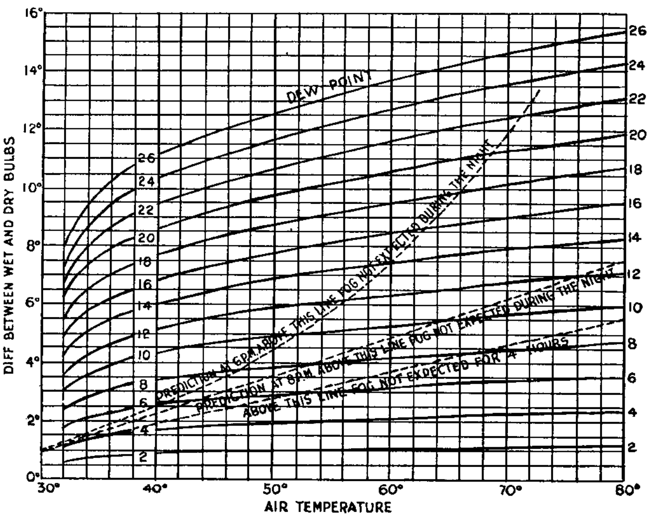

Fig. 6. Taylor's fog prediction diagram for Kew on calm, clear nights. The statistical «fog prediction lines» may conveniently be plotted on Fig. 5.

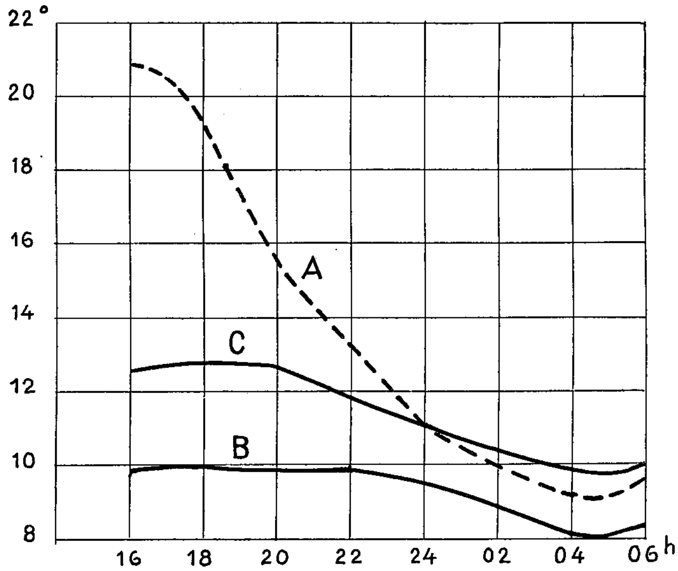

Fig. 7. Curve *A* shows the normal fall in temperature on calm and clear nights at Kew. Curve *B* shows the normal variation in dew-point on calm and clear nights without fog; and curve *C* shows the normal variation in dew-point when fog developed in the night.

calm evenings, but also the normal variation in dew-point. It is well known that the dew-point temperature has a diurnal variation which is slight as compared with that of the air temperature. Fig. 7 shows this. It will be noticed that the dew-point temperature normally remains constant during the day, but a slight fall sets in after sunset on account of dew that forms on the ground. When a fog forms, the dew-point decreases still further, owing to the condensation of water vapour. In the morning, the dew-point rises again, as the dew, or the fog, evaporates.

If aerological ascents are available, one may go a step further and try to estimate the diurnal heating and cooling by considering the actual stability conditions of the air. It follows from the theory of mixing that the more stable the air the greater is the diurnal amplitude of temperature. Similarly, the presence of an inversion above the ground has a marked influence on the magnitude of the diurnal variation[1]).

The above refers mainly to radiation fogs, but similar reasonings hold for all types of fogs (except steam fogs) since it is immatrial whether the cooling is caused by radiation or not. However,

[1]) *Petterssen*, S., On the Causes and the Forecasting of the California Fog. Bul. of the Am. Met. Soc. Vol. 19, No. 2, 1938, pp. 49—55.

in predicting non-radiation fogs, one must estimate the probable changes of temperature along the trajectory of the air. The knowledge of average local conditions (such as are illustrated in Fig. 6) is then less useful.

Radiation fogs are always connected with an inversion of temperature and specific humidity at the ground. The same is true, in most cases, of advection fogs. If the wind is strong enough to raise the inversion, the fog dissipates, or develops into a layer of stratus. In rare cases, however, advection fogs may persist even when the inversion is at some considerable distance above the ground. This happens when the air travels over a surface which is so much colder than the air that the cooling from below overcompensates the effect of vertical mixing. The same also applies to fogs that form through the combined influence of advection and evaporation from rain that falls from warm air aloft (refer to Fig. 1 *C* and *D*). Fogs that exist together with strong winds are usually very deep, because they fill the entire layer below the inversion which has been raised by turbulent mixing.

8. **Fog in Relation to Snow.** As cooling from the underlying surface is the primary cause of formation of fog, it is surprising to observe how rarely fogs develop over cold continents, such as Siberia and Canada, in winter. It is frequently observed that moist air from the Atlantic Ocean enters the snow-covered regions of North Europe and Siberia without producing fog. On the other hand, when Atlantic air comes over the Labrador Current, it produces fog very quickly. In the first case the air may be cooled from 15° C to — 40° C; and in the latter case it may be cooled only from 15° C to 5° C. Even when the air is perfectly calm and clear, fogs do not easily form over cold snow-fields. The principal reason for this may be sought in the depression of the saturation vapour pressure over ice.

In Fig. 8, the curve ΔE represents the difference in saturation pressure over water and over ice. It is zero at 0° C, and it has a maximum (0.27 mb.) when the air temperature is about — 12° C, and it decreases as the air temperature decreases still further. The curve R represents the relative humidity of the air when the vapour pressure of the air equals the saturation pressure over ice. It will be seen that R decreases linearly with derceasing temperture.

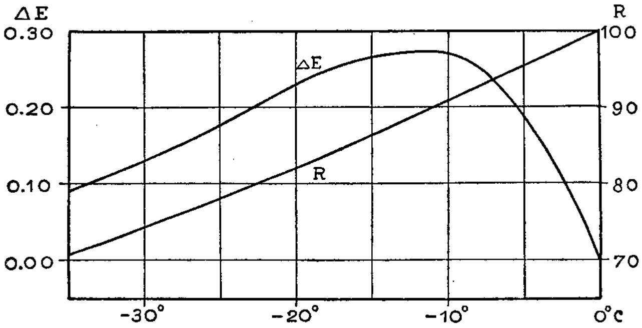

Fig. 8. Difference in saturation pressure over water and over ice (curve ΔE); and relative humidity (curve R) of air which is saturated over ice.

When air over a snow covered surface is cooled to such an extent that its relative humidity increases to the value indicated by R in Fig. 8, water vapour is sublimated on the snowy surface. As this continues, the air obtains a minimum of specific humidity near the surface. The eddy flux of moisture will then be directed downwards, and the air column gives off moisture to the snow. When the air temperature is slightly below zero, ΔE is small, and if the air is cooled with a sufficient speed, a fog may easily form. But the fog must then give off moisture to the snow. The fog can only persist when the air is steadily cooled (e. g. by advection), or when the downward eddy flux of moisture is strong enough to overcompensate the loss of moisture caused by sublimation on the snow.

As the air temperature decreases and approaches — 10 to — 15° C. ΔE increases to its maximum value, and the drying influence increases accordingly: it becomes increasingly difficult for a fog to form; and a fog that is brought over a cold, snow-covered surface has a marked tendency to dissipate.

As the diurnal amplitude is very small when the air temperature is very low, a fog will not form over a cold snow-covered surface as a result of nocturnal cooling only. The fogs that are sometimes observed over cold continents in winter, are caused by advectien of moist air from warmer regions. When such air is cooled rapidly, it may produce fog which gradually dissipates as the air temperature decreases. As was shown in the last paragraph, the «mist interval» is very wide at low temperatures. Most of the fogs that occur over cold snow fields are very light; they are often so shallow that the sky is discernible.

The above conditions apply when the air temperature above the snow is below 0° C. The conditions are, however, different when the air temperature above the snow is higher than 0° C. The snow is then melting. Under such conditions, as has been shown by *Sverdrup*[1]), the air that is in direct contact with the melting snow has a temperature which is close to 0° C, and a relative humidity of 100 percent. The vapour pressure of the air which is in contact with the melting snow is then 6.11 millibar. If the spesific humidity increases along the vertical (as it is likely to do when the air temperature increases along the vertical), there is an eddy flux of moisture downwards, and moisture from the air is condensed on the snow. This implies that condensation occurs on the melting snow whenever the *relative humidity* of the air exceeds a limiting value which depends on the air temperature above the snow. These limiting values are given in Table 4. If the relative humidity for any given air temperature is less than the limiting value, water evaporates from the snow. *Sverdrup* (loc. cit.) points out that relative humidities less than the limiting values are seldom encountered over melting snow, and as a result, condensation of water vapour is the normal occurrence in spring after the air temperature has become greather than 0° C. The observations of *Ahlmann* and *Sverdrup* show that water is condensed on the melting snow in considerable amounts and that the latent heat thereby liberated, results in appreciable ablation.

Table 4.

Air temperature at some distance above the ground (°C)	Saturation vapour pressure (millibars)	Limiting value of relative humidity (percent)
0	6.11	100
2	7.05	87
4	8.13	77
6	9.35	67
8	10.73	58
10	12.28	51

[1]) *Sverdrup*, H. U. The Ablation on Isachsen's Plateau and on the Fourteenth of July Glacier in Relation to Radiation and Meteorological Conditions. Scientific Results of the Norwegian-Swedish Spitsbergen Expedition in 1934 Part IV. Geografiska Annaler, 1936.

The above results are of considerable interest in connection with our discussion of formation and dissipation of fog over snow. Suppose that an air temperature of 4° C is measured at a height of 2 metres above the snow. If the air were saturated, there would be a vapour pressure gradient of 1 mb. per metre in the layer below the thermometer. Water vapour would then condense quickly on the snow which would retain a temperature of 0° C. The downward flux of moisture would continue until the relative humidity has been lowered to the limiting value 77 percent (see Table 4). In order to maintain saturation, the advection of warmer air would have to be strong enough to overcompensate the loss of moisture towards the snow. The higher the air temperature above the snow, the more difficult would it be for saturation (or fog) to be maintained. It follows then that fogs can form only in exceptional cases when the air temperature at some distance above the ground is much above 0° C. Similarly, fogs that are transported over melting snow will have a marked tendency to dissipate. From this we may conclude *that fogs over melting snow can exist only when the air temperature is not far removed from 0° C.*

When the air temperature falls below 0° C, the dissipating influence due to the depression of the saturation water vapour pressure over snow commences, and increases as the temperature decreases. It follows then that the highest frequency of fog over snow-covered ground in spring would be observed when the air temperature is close to 0° C.

The arctic fogs, which are so frequent in summer, form because of advection of warm air over cold sea. Such fogs often invade the arctic fields of ice, where they gradually dissipate, unless the advection of moist air from the open sea is sufficiently intense to overcompensate the effect of condensation on the melting snow. The arctic summer fogs are, therefore, essentially advection fogs.

The dissipating influence of snow on fogs is brought to light by statistical evidence. An example is shown in Table 5. Let f denote the frequency of occurrence of fog in the various temperature intervals, and let F be the frequency of air temperatures observed within these intervals. Table 5 then shows the percentage frequency ($100\ f/F$) of occurrence of fog for any given temperature in Oslo.

Table 5.

Fog Frequencies (100 f/F), Oslo 1920—1931.

(Total number of observations 4365.)

Temperature intervals (°C)	October	Jan.–Febr.	March
20 – 15	0.0	—	0.0
15 – 10	0.5	—	0.0
10 – 5	3.0	0.0	0.0
5 – 0	5.7	5.6	3.3
0 – – 5	12.3	5.0	8.7
– 5 – – 10	0.0[1])	4.6	6.0
– 10 – – 15	—	4.1	0.0
– 15 – – 20	—	0.0	—
– 20 – – 25	—	0.0	—

It is of interest to note that October in Oslo is a snowless month, while January and February normally have snow with freezing temperatures, and March frequently has snow with temperatures above freezing (melting snow)[2]). It will be seen from the above table that the frequency of fog in October is highest when the air temperature is lowest. This is what one would expect over snowless ground: The lower the temperature sinks the greater is, on the average, the probability of formation of fog. This particularly applies to radiation fogs.

In January and February, when the ground is covered by snow, the conditions are quite different. The highest frequency of fog over snow occurs when the temperature is about 0° C, while, for lower temperatures, the fog frequency decreaces. The circumstance that the highest frequency of fog occurs at high temperatures in mid-winter seems to indicate that the mid-winter fogs in Oslo are mainly advection fogs which form when there is a transport of warm and moist air over the cold surface. The fogs then persist as long as the advection is strong enough to overcompensate the dissipating influence of the snow. As the air temperature approaches — 10° C to — 15° C, the difference in saturation pressure over water and ice increases to its maximum value (0.27 mb), and sublimation on the snow increases accordingly.

[1]) There are only 3 temperature observations in this interval.

[2]) In the period examined, the average number of days with ground completely covered by snow is as follows: October, 0.6; January, 21.3; February 21.8; March 14.6.

In March, when the mean air temperature is above freezing, the snow is normally melting. Fogs with air temperature considerably above 0° C are then very rare. The highest frequency occurs at temperatures just below zero, and the frequency decreases as the temperature decreases further. The spring fogs in Oslo are apparently also mainly of the advection type, and it is plausible that advection is the allimportant cause of formation of fogs over snow-covered ground.

What has been said above applies to *water fogs*. At low temperatures, the water vapour of the air sublimates on the sublimation nuclei, and *ice crystal fogs* result. Such fogs are sometimes observed at temperatures about — 10° C, but, most frequently, sublimation begins at about — 20° C. At still lower temperatures (— 30° C to — 50° C) ice crystal fogs are the normal occurrence (e. g. in Siberia and North Canada in winter). At such low temperatures, the specific humidity of the air is insufficient for producing a water cloud of such density that the horizontal range of visibility is reduced to 1000 metres or less. However, when sublimation occurs, a much smaller amount of aqueous vapour is needed in order to produce ice crystals whose size and number suffice to reduce the horizontal visibility below the fog limit. At extremely low temperatures, the density of the cloud is such that it appears as a thin mist or haze («frost haze»).

It is of interest to remark that *snow on the ground has no dissipating influence on ice crystal fogs*. It is, therefore, to be expected that the fog frequency would show an increase in the temperature intervals below — 20° C, or so, where ice crystal fogs are predominant. In order to test this assumption, fog frequencies have been computed for 10 synoptic stations in the Siberian lowland east of the Ural Mountains. The result is shown in Table 6, which is computed as explained above

It will be seen that the fog frequency, here too, decreases rapidly as the air temperature sinks below 0° C. The decrease continues until the air temperature approaches — 20° C, where there is a hump, which perhaps is due to the circumstance that water fogs as well as ice crystal fogs may occur within this temperature interval. The frequency then decreases slightly, and increases again as the temperature decreases to — 50° C. At still lower temperatures, the frequency would be likely

Table 6.

Fog Frequency (100 f/F) at 10 Siberian Lowland Stations, October to March.

(Total number of observations 4389.)

Temp.intervals (°C)	15/11	10/6	5/1	0/–4	–5/–9	–10/–14	–15/–19
Fog frequency (%)	0.0[1])	9.0[1])	9.1	4.9	4.3	2.7	2.7
Temp.intervals (°C)...........	–20/–24	–25/–29	–30/–34	–35/–39	–40/–44	–45/–49	–50/–55
Fog frequency (%)	7.0	5.0	4.4	6.5	12.8	15.4	0.0[2])

to decrease, on account of the slight amount of specific humidity which the air can hold at such low temperatures.

Summing up, we may say that melting snow has a marked dissipating influence on fog, when the air temperature is above freezing. The dissipating influence is greater the higher the air temperature rises above 0° C. As the air temperature falls to 0° C, this dissipating influence vanishes; but when the air temperature sinks below 0° C, another dissipating influence, due to the depression of the saturation pressure over ice, commences, and increases in intensity until the air temperature reaches — 10° C to — 15° C, where it has a maximum. This dissipating influence is identical with the «ice crystal effect» met with in the upper portion of clouds, where ice crystals and water droplets coexist. *Bergeron*[3]) ascribes the release of precipitation from clouds to this effect. As the air temperature sinks to approximately — 20° C, sublimation nuclei become active, and ice crystal fogs form. The presence of snow on the ground has then no dissipating influence on the fogs.

It is of interest to remark that the dissipating influences mentioned above are not sufficiently strong completely to prevent the formation of fog. However, when they are present, a corresponding fog producing process (e. g. advection) is necessary for its production and maintenance. It should also be mentioned that the deposition of water from the air on the snowy surface is of considerable importance in the production of polar continental air masses in winter[1]).

9. Classification of Fogs. In the above paragraphs an attempt has been made to explain the physical processes which, under favourable conditions, may lead to the formation of fogs. Mention has also been made of the processes which counteract the fog producing agencies, and which may lead to the dissipation of existing fogs. Based on this discussion, fogs may be classified according to the identity of the process which constitutes the principal cause of their formation. This is shown in the left hand portion of Table 7. A summary of the fog dissipating agencies is given in the right hand portion of the same table.

In many cases, it will be found that fogs are produced as a result of several of the above fog creating processes, and these are often counteracted by one or more of the fog dissipating processes. The tropical west-coast fogs[2]), perhaps, offer the best example of the multitude of factors which may be involved in the production of fogs. From the point of view of weather forecasting, a detailed discussion of the various combinations of fog producing and fog dissipating influences is of great interest. These questions have been discussed by *Willet*[3]) and *Petterssen*[4]).

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[1]) Probably no snow on the ground.

[2]) There is only one temperature observation in this interval.

[3]) *Bergeron*, T., On the Physics of Clouds. Mémoire présenté a l'Association de Météorologie de l'U. G. G. I. Lisbon, 1933.

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[1]) *Willet*, H. C., American Air Mass Properties. Papers in in Physical Oceanography and Meteorology, Cambridge, Mass. 1933.

[2]) *Petterssen*, S., loc. cit. p. 17.

[3]) *Willet*, H. C., Fog and Haze, etc., Monthly Weather Review, vol. 56, 1928, pp. 435—468.

[4]) *Petterssen*, S., loc. cit. p. 10.

Table 7.
Classification of Fog.

Fog Producing Processes	Fog Dissipating Processes
I. *Evaporation* from: 1) Rain which is warmer than the air (rain area fog, or *frontal fog*). 2) Water surface which is warmer than the air (*steam fog*).	I. *Sublimation or Condensation* on: 1) Snow with air temperature below 0° C (excepting ice crystal fogs). 2) Snow with air temperature above 0° C (melting snow).
II. *Cooling due* to: 3) Adiabatic upslope motion (*upslope fog*). 4) Flux of air across the isobars towards lower pressure («*isobaric fog*»; effect negligible). 5) Falling pressure («*isallobaric fog*»; unimportant). 6) Radiation from the underlying surface (*radiation fog*). 7) Advection of warmer air over a colder surface (*advection fog*).	II. *Heating* due to: 3) Adiabatic downslope motion. 4) Flux of air across the isobars towards higher pressure (effect negligible). 5) Rising pressure (unimportant). 6) Radiation absorbed by the fog or by the underlying surface. 7) Advection of colder air over a warmer surface.
III. *Mixing.* 8) Horizontal mixing (unimportant by itself and strongly counteracted by vertical mixing).	III. *Mixing.* 8) Vertical mixing (important in dissipating of fogs and producing stratus).

Printed June 19, 1939.